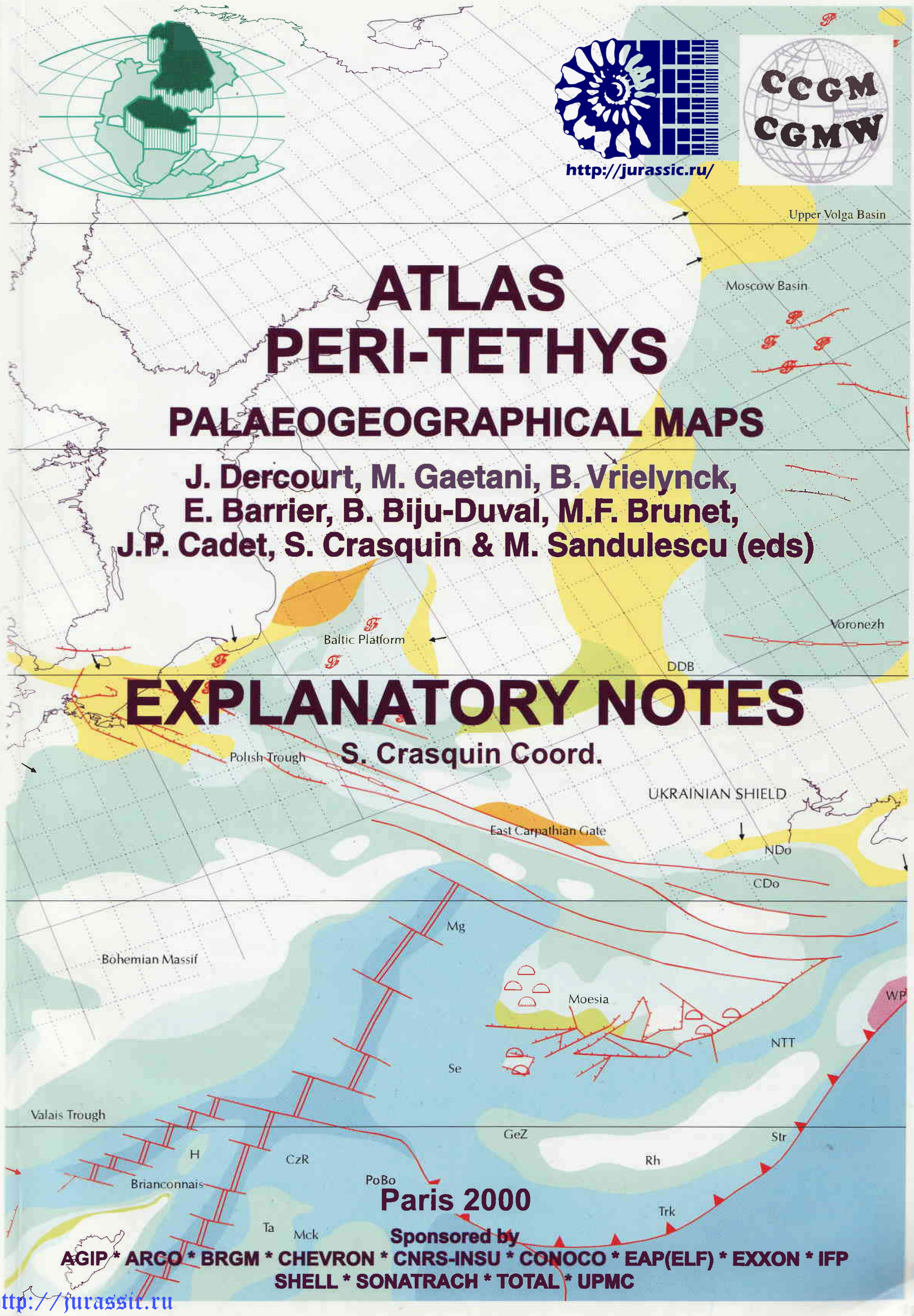

ATLAS PERI-TETHYS

PALAEOGEOGRAPHICAL MAPS

J. Dercourt, M. Gaetani, B. Vrielynck,
E. Barrier, B. Biju-Duval, M.F. Brunet,
J.P. Cadet, S. Crasquin & M. Sandulescu (eds)

EXPLANATORY NOTES

S. Crasquin Coord.

Paris 2000

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AGIP * ARCO * BRGM * CHEVRON * CNRS-INSU * CONOCO * EAP(ELF) * EXXON * IFP
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PERI-TETHYS
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The Atlas, the maps of the Atlas and the explanatory notes should be cited as follow:

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Maps of the Atlas:

THIERRY J. *et al.* (41 co-authors) (2000).– Early Tithonian. *In*: DERCOURT J., GAETANI M. *et al.* (eds), Atlas Peri-Tethys, Palaeogeographical maps. *CCGM/CGMW*, Paris: map 11.

Explanatory notes:

THIERRY J. (2000).– Early Tithonian. *In*: S. CRASQUIN (coord.), Atlas Peri-Tethys, Palaeogeographical maps - Explanatory notes. *CCGM/CGMW*, Paris: 99-110.

Nous dédions cet Atlas à la mémoire de
M. A. Boudjema,
représentant de la SONATRACH,
assassiné à Alger en 1994

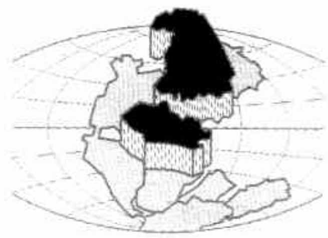
We dedicate this Atlas to the memory of
Mr. A. Boudjema,
SONATRACH representative,
murdered in Algiers in 1994

				(1)	(2)	Map number
IV		Holocene				
		Pleistocene		1.75	1.81	24
CAINOZOIC	NEOGENE	Pliocene	Gelasian		2.58	
			Piacenzian	3.4	3.60	23
			Zancleian	5.3	5.33	
		Miocene	Messinian	7.3	7.1	22
			Tortonian	11.0	11	
			Serravalian	14.3	13.6	21
			Langhian	15.8	16.4	
			Burdigalian	20.3	19.1	20
			Aquitanian	23.5	23.8	
	PALAEOGENE	Oligocene	Chattian	28		19
			Rupelian	33.7		
		Eocene	Priabonian	37.0		
			Bartonian	40		18
			Lutetian	46.0		
			Ypresian	53		17
		Palaeocene	Thanetian			
			Selandian			
			Danian	65.0	65.0	16
MESOZOIC	CRETACEOUS	Late	Maastrichtian	72.0	71.3	15
			Campanian	83	83.5	
			Santonian	87	85.8	
			Coniacian	88	89.0	
			Turonian	92	93.5	14
			Cenomanian	96	98.9	
		Early	Albian	108	112.2	13
			Aptian	113	121.0	
			Barremian	117	127	12
			Hauterivian	123	132.0	
			Valanginian	131	136.5	
			Berriasian	135	144.2	
	JURASSIC	Late	Tithonian	141		11
			Kimmeridgian	146		10
			Oxfordian	154		
		Middle	Callovian	160		9
			Bathonian	164		
			Bajocian	170		
			Aalenian	175		8
		Early	Toarcian	184		
			Pliensbachian	191		
			Sinemurian	200		7
			Hettangian	203		
	TRIASSIC	Late	Rhaetian			
			Norian	220		6
			Carnian	230		5
		Middle	Ladinian	233		
			Anisian	240		
		Early	Olenekian			4
			Induan	250	251.1	

				(1)	(2)	Map number
PALAEOZOIC	PERMIAN	Lopingian	Changhsingian	250	251.1	
			Wuchiapigian			
		Guada-lupian	Capitanian			
			Wordian			3
			Roadian		272.2	
		Cisuralian	Kungurian			
			Artinskian		280.3	2
			Sakmarian			
			Asselian	295	298	
	CARBONIFEROUS	Pennsyl-vanian	Gzhelian			
			Kazimovian			
			Moscovian			1
			Bashkirian		320	
		Missis-sippian	Serpukovian	345	342	
			Visean			
			Tournaisian	355	354	

(1) after Odin (1994)
(2) Subcommission of ISC *in* International Stratigraphic Chart (2000)

Position of the Atlas Peri-Tethys maps



n°	Map	age
1	Moscovian	312-305 Ma
2	Artinskian	280-273 Ma
3	Wordian	266-264 Ma
4	Olenekian	245-243 Ma
5	Early Ladinian	238-235 Ma
6	Late Norian	215-212 Ma
7	Late Sinemurian	193-191 Ma
8	Middle Toarcian	180-178 Ma
9	Middle Callovian	157-155 Ma
10	Early Kimmeridgian	146-144 Ma
11	Early Tithonian	141-139 Ma
12	Early Hauterivian	123-121 Ma
13	Early Aptian	114-112 Ma
14	Late Cenomanian	94.7-93.5 Ma
15	Early Campanian	83-80.5 Ma
16	Late Maastrichtian	69.5-65 Ma
17	Early - Middle Ypresian	55-51 Ma
18	Late Lutetian	44-41 Ma
19	Late Rupelian	32-29 Ma
20	Early Burdigalian	20.5-19 Ma
21	Early Langhian	16.4-15.5 Ma
22	Late Tortonian	8-7 Ma
23	Piacenzian/Gelasian	3.4-1.8 Ma
24	Last Glacial Maximum	20-18 Ka

Map leaders

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Wordian to Norian: **Maurizio Gaetani**, Milano University, Italy

Sinemurian to Tithonian: **Jacques Thierry**, Bourgogne University, Dijon, France

Hauterivian to Maastrichtian: **Jean Philip**, Aix-Marseille University, France

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Foreword

The 24 maps (1:10.000.000) of the Peri-Tethys Programme are the result of a very wide effort of hundreds of scientists from Europe, Africa and Asia. The process to produce the maps started from the palaeoenvironments and the palaeogeography, that were under the responsibility of 6 map-leaders, G.B. VAI (Moscovian - Artinskian), M. GAETANI (Wordian to Norian), J. THIERRY (Sinemurian to Tithonian), J. PHILIP (Hauterivian to Maastrichtian), J. MEULEMKAMP (Ypresian to Piacenzian) and J-P. PEULVAST (Pleistocene). Each map leader organized regional working-groups and in some instances he decided to share the responsibility for single maps with others. The map-leader discussed the results of the regional working groups with the involved scientists in a large number of meetings, held in many localities. The first draft that resulted from this phase of collecting data was discussed and improved with regional experts, that provided the regional cartographic material. Often, the regional experts to provide the cartographic material used data supplied by other colleagues, that were not actually involved in the cartographic process, but gave advice, data and information. As results of this process, the map leader(s) prepared the final version of the map, trying to unifying in a coherent view the facies pattern. The map leader(s) are labelled on the map as the author(s); the scientists who provided the regional cartographic documents are labelled as co-authors and the area they supplied is indicated in the index- map with figures; the scientists who supplied information, data and advice are labelled as contributors. A similar process was undertaken for the structural part, which was leaded by E. BARRIER & J.P. CADET for all the maps. They added to the maps the tectonic data, with the help of regional experts, on the basis of the already produced draft for palaeoenvironments and palaeogeography, in some case also supplied with some palaeotectonic information. They appears as authors and the regional experts for the structures appears as co-authors, labelled with letters on the index-map. Additional data for the subsiding basins were supplied by scientists involved in the modelling sub-project, as well as the palaeostress data by the palaeostress people.

The Explanatory Notes has been largely written by the author(s) of the maps. They summarized the results of the regional working groups and/or used for particular areas the text supplied by some of the co-authors. Also in this case, the author(s) made an effort to homogenise the data and to present them under a coherent view.

The drawing of the maps has been done mostly at the scale 1:5.000.000 that were provided to the map-leaders by B. VRIELYNCK, who managed all the informatic process of the project. Since it was decided to start from the geodynamic reconstructions of the Tethys Project (DERCOURT *et al.*, 1993), the first step was to transform the oblique rectilinear projection in rectilinear projection. For the ten maps which were not yet drawn in the Tethys Project, the geodynamic reconstructions has been done taking into account, at this step, the kinematic parameters proposed by L.E. RICOU (1993, 1996).

The first hand drafts of the environment, and some tectonic features of the Moscovian to Norian maps were digitalized in Milano by Gl. NOZZA with the Arc-Info software. Jurassic maps were digitalized by BRGM with Intergraph Microstation software. Cretaceous maps were digitalised by the Geographica office with Adobe Illustrator software. The Cainozoic maps were digitalized in Utrecht by T. VAN HINTE with Autocad software. Sent to Paris, these digitalized maps have been imported in Arc-Info. Then, to the spatial data included by the map-leaders, the Tethys ones have been added at their turn.

Arc/Info software is a Geographic Information System. It combines spatial features and factual data relative to the spatial ones. Factual data, recorded as tabular data, compose a database of which structure and management were supervised by B. VRIELYNCK. For each map seven main tables are included: environment (35 sedimentary domains are distinguished), lithology, (78 facies have been depicted), marine currents (4 types are recorded), tectonics (11 features are distinguished), palaeostress, palaeolatitudes and actual coasts.

After a number of revisions and improvements, the palaeoenvironments maps were completed with the tectonic features supplied by the tectonic team. The largest part of this work has been done in Paris. Then, the final version was produced and approved by the palaeoenvironment and tectonic teams and sent to the BRGM for printing.

One major scientific problem that arose during the progress of the map production was how to draft the Tethyan part. The Tethys project consisted of 14 maps, the present Atlas include 24 maps. Beyond the Pleistocene that has no palaeogeographic and geodynamic problems, 9 maps has no counterpart in the Tethys project, and for some of the others, the palaeogeography was thought to need up-grading. We solved the problem making some upgrading in the Tethys part for some maps, in other maps tracing only major features, but not details. We also avoided to print the lithologies in the Tethyan area, but only the general facies pattern. This policy was decided because in was not in the present project to fully reconsider the Tethyan part. However, it is obvious that almost 10 years of researches and the much more detailed scale of the Peri-Tethys programme give evidence for the necessity of a general reconsideration for the Western Tethys evolution and palaeogeography.

The tectonic evolution of the platforms surrounding the Tethyan domain was one of the initial themes of the Peri-Tethys Programme. The major objective of the tectonic group of PTP was to provide a simple synthetic view of the tectonic activity for each time-slice corresponding to the 22 maps between Late

Permian and Quaternary. The main final product of the Peri-Tethys Programme, an atlas of palinspastic maps, imposes several important constrains to the tectonic data incorporated to the maps. They have been devised (1) to portrait one particular tectonic context on the maps, and (2) to complement the associated palaeoenvironmental data constituting the base of the maps.

The choice of periods for all the maps exclusively followed stratigraphic criteria. The time-slices generally correspond to well defined biostratigraphic intervals. They range between 1 and 4.5 Ma, with a mean time-slice of 2 Ma. These periods are short compared to the duration of most of the major tectonic events that may last for several tens of million years. This accuracy depends from the type of tectonic event (extensional, compressional), the regional stratigraphy, and the available data (field and/or subsurface data). As one single tectonic event commonly lasts several millions of years for the Mesozoic and Tertiary maps the duration of the time-slices generally allows to integrate one tectonic event (or a lack of tectonics) in one region for a given map. On the other hand, the mean time interval between two maps (from Wordian to Piacenzian) is of 14 Ma. As a consequence, short tectonic events, such as minor rifting or inversions, that lasted only few million years, do not necessarily appear on the maps. So, from a tectonic point of view each map must be considered as a flash on the tectonic history of the Peri-Tethyan domain.

Most of the tectonic works focus on regional topics such as the studies of orogenies, the evolution of basins, the palaeostress evolution of platforms, or particular rifting. Because of this regional approach, in a first step, and before to begin the mapping of the tectonic data, we prepared regional syntheses on the tectonic evolution of the main regions of the Peri-Tethyan domain. These syntheses were elaborated together with regional specialists representing different domains of tectonics (structural geology, palaeostress, basin analysis, modelling) and summarised in tectonic logs. For each tectonic event, the type of tectonics, the age, and the main stress pattern have been defined. A special attention was paid to the uncertainty on the ages of the tectonic events. These syntheses concern both the northern and southern Tethyan margins as well as some interesting and well studied regions of the Tethyan domain. The second step was devoted to the representation of the tectonic events on the maps.

Basically, three main types of data were reported on the map: the major structures, the palaeostresses, and the axes of subsidence of the basins. The basic idea was to restore the tectonic context integrating these three kinds of independent data into the maps. Among them, the major tectonic features provide the best image of the tectonics (type and geographical extent). The most frequent tectonic contexts represented on the maps are:

- the rifting periods - the main normal faults are mapped as well as the subsidence axes; the post-rift passive thermal subsidence is represented by thermal subsidence axes;
- the orogenies - the major thrusts and anticline and/or syncline axes are reported. The largest subsiding fore deep basins in front of the major active orogens appear as flexural axes;
- simply inverted basins are figured like the orogenic belts. They are only differentiated by the size of the deformation zone and the density of active tectonic features;
- the major transcurrent fault zones are only indicated as strike-slip faults with the sense of displacement. The eventual pull-apart basins associated to these strike-slip faults are never figured; they are too small with respect to the scale of the maps.

The palaeostress data appear in all the previously defined tectonic contexts as arrows differentiating extensional, reverse and strike-slip regimes. Each of these synthetic arrows represent a population of homogeneous palaeostress tensors reconstructed in a small area from the analysis of fault-slip data sets. They define mean regional Palaeostress regimes synthesised from a Palaeo-stress database compiled from the published data and from original works supported by the Peri-Tethys Programme.

M. GAETANI, B. VRIELYNCK & E. BARRIER

GENERAL FEATURES OF THE ATLAS

J. DERCOURT[1], M. GAETANI[2] & B. VRIELYNCK[3]

I.- INTRODUCTION

The set of the 24 maps, from Moscovian to latest Pleistocene, encompasses a time span of 300 Ma, and covers the Peri-Tethyan regions from the Atlantic to the Urals and Aral sea to the north, and from the Atlantic to the Gulf to the south. Many geodynamic events took place in such long span of time. The sea-ways of Tethys evolved between these two broad regions and their behaviour is believed to have strongly affected the evolution of the Peri-Tethyan areas. Of major importance were also the final stages of the Hercynian orogeny and the opening of Central Atlantic ocean.

The whole period could be subdivided in three major intervals, that we might define as the Pangaea time, the Tethys time and the Alpine time. The Pangaea time includes the interval from the Moscovian to the beginning of the Late Triassic, the Tethys time the interval from latest Triassic to the mid Cretaceous, and the Alpine time the more recent interval.

II.- DIFFERENT TIMES

II.1.- The Pangaea times (map 1 - Moscovian to map 5 - Early Ladinian)

This interval, about 100 Ma long is mostly a time of convergence, important lateral displacements and oblique rifting. A number of basic features may be recognised, like (Fig. I):

- the final assembling of the Pangaea, including the closure of the Ural sea-way and the final building of the Ural Mountain Range
- the ongoing subduction of the Palaeo-Tethys below the Eurasian margin with the formation of the Altaid Orogenic system and then the consolidation of the Turan block
- the opening of the Neo-Tethys with the transit of the Peri-Gondwanan fringe blocks from Gondwana towards Eurasia and their final welding against the northern continent
- the large lateral displacements between Africa and Eurasia, with local transtensions, pull-apart basin opening and transpressions.

Resulting from these major kinematic trends, the Peri-Tethyan regions were largely under continental conditions during the Pangaea time, independently from the dominating glacio-eustatic low-standing during the Late Carboniferous and earliest Permian, which is only marginally reflected in the Moscovian map. Moscovian was an interval a relative high-standing of the sea-level.

II.1.1.- Palaeoenvironments

A large variety of environments is shown on the maps, due to the significant climatic changes, relief formation and erosion during the Pangaea time.

Because of the Hercynian orogeny in Europe, the formation of Mauretanides and Appalachians and of the Ural - Altaid orogenic system, a number of mountain ranges bounded the Peri-Tethyan regions or were elevated within the Peri-Tethyan regions, supplying and feeding a large amount of debris to the foredeep, piggy back or back arc basins. Fluviatile and lacustrine deposits are consequently very widespread. When the coastal relief was or became more gentle because of progressive flattening during the Permian and Triassic, important interfingering with the marginal brackish seas developed, over very wide flats. The most typical is the flat of Arabia towards Neo-Tethys, which was about 1000 km wide. Along the shores of the Tethys, epicontinental seas extended over not very deep floors, sometimes as wide

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as 1000 km like the Mid European Basin and the Peri-Caspian Basin during the Middle Triassic.

The climate evolved from the severe gradients of the Moscovian, where ever wet forests flourished in the Eurasian belt, and glaciogenic deposits are observed in Oman, to the increasingly arid weather of the Permian and eventually to the arid/semiarid climate of the Triassic. The continental basins were consequently filled also by playa deposits. In particular conditions, often in time slices not represented by the maps, huge salt deposits, periodically fed by the neighbouring shallow seas, filled up the depression. The Kungurian salts of the Precaspian Depression, the post-Wordian Zechstein in the Polish Trough, German and North Sea basins, the Bellerophon basin in the Southern Alps and the Khuff Member B to D in the Arabic peninsula are examples of such enormous subtractions of salts to the sea water.

In the epicontinental seas, algal and coral assemblages, were able to build up significant carbonate banks, in which fusulinids, brachiopods and other invertebrates dwelled. After the Permo-Triassic crisis, since the Middle Triassic, the "carbonate factory" was fully at work. Most of the marginal seas were in low-temperate to equatorial conditions, consequently carbonate banks and ramps are wide spread. In the Norian, peculiar geochemical conditions of the sea-waters allowed to formation of the extremely wide carbonate platform from Betic margin to the Peri-Gondwanan Fringe and Oman which was subsequently penecontemporaneously dolomitised.

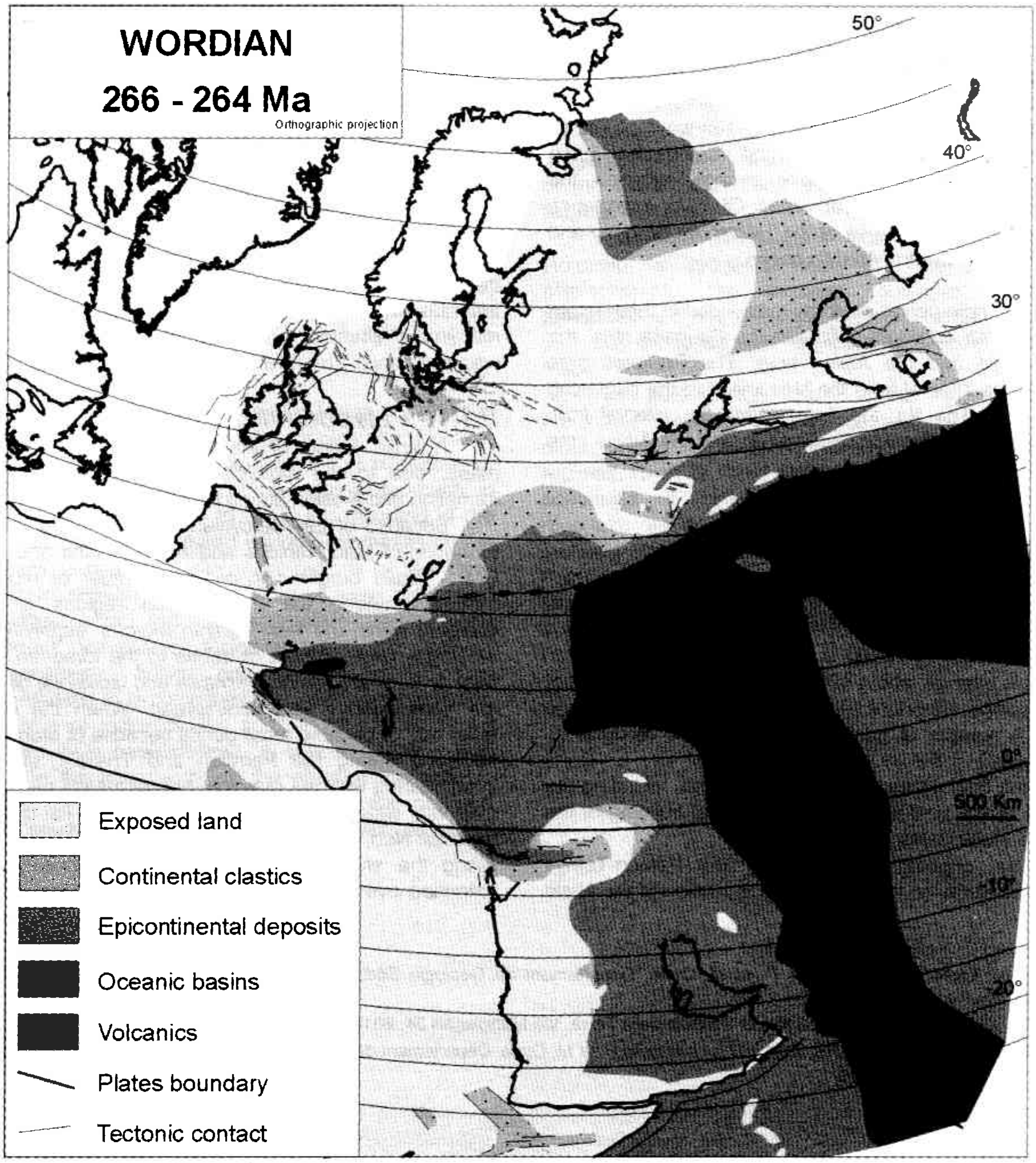

Fig. I: Latitudinal zonation (Wordian).

Note (i) the complete absence of Atlantic basin, (ii) the importance of Cimmerian Transit Plate, and (iii) the subsidence of Northern Peri-Hercynian basin (Polish Trough along the Teisseyre - Tornquist Line).

II.1.2.- Structural evolution

The kinematics of the Pangaea time are made complex by the interplay of three main trends.

To the east, the continuous activity of the Transit Plate (RICOU, 1993, 1996), was the driving mechanism to subduct the Turkestan oceanic lithosphere during the Carboniferous and Early Permian and, later, the Palaeo-Tethys oceanic lithosphere below the Asian margin. The rotation of the Transit Plate was also the trigging mechanism to open the Neo-Tethys Ocean during the Permian, between the Peri-Gondwanan fringe blocks and the India and the Arabia parts of Gondwana. The continental blocks involved were progressively accreted to the Asian margin, originating volcanic arcs and back arc basins in the Caucasus - Caspian - Turan regions.

The north-westward rotation of Gondwana against Laurussia with the eventually progressively hetero-chronous collision during the Carboniferous and the Early Permian, led amongst others to the formations of the Mauretanides and the Appalachians. The lateral displacement involved was dextral. Most intriguing is the time span from Middle Permian to Middle Triassic, roughly from 270 to 230 Ma. During this interval the convergence was gradually over and thermal and gravity relaxation affected the mountains areas, with local rifting, transtension and transpression. In this situation the Neo-Tethys showed trends to propagate westwards, dissecting part of the convergent zone in the future southern Europe. The results is a complicate palaeogeo-graphic pattern, still not properly understood. This trend ended in the Carnian (Late Triassic), when progressively propagating rifting from the future central Atlantic Ocean, inverted it to a sinistral movement. The pivot time, Carnian, is very well constrained, both on the Moroccan side both on the North American side eventually leading to the onset of the great Central Atlantic Magmatic Province. On a broader perspective, this is a part of the big displacement leading from the Pangaea B to the Pangaea A configurations.

The North-West Europe, after the collision and building of the Hercynian orogen, underwent a horst and graben evolution, partly with a significant N-S component, and partly with a NW-SE component. This is the result of the orogenic collapse following the acme of the convergence, of reflex of the rifting trend active in the area of future North Atlantic, and of the different displacement speeds between the Baltic Shield and East European Platform from one side and the Western-Central Europe from the other side, along the Tornquist-Teisseyre alignment.

II.1.3.- Magmatism

During the Pangaea time, magmatism was obviously very spread in the orogens. But also outside the main orogenic belts, volcanic rocks are frequent and largely of calcoalcaline signature. In the Turan, Altaid and Pre-Uralian depression, volcanoclastic products are mainly preserved and their composition varies from andesitic to dacitic and ryolitic.

In the Caucasus, Dobrogea and the whole western Europe bimodal volcanics are particularly spread during the Permian, linked to the transtensive and transpressive lineaments. Bimodal volcanism is also present during the Triassic, but it is more concentrated within the branches of the Tethys, than on the Peri-Tethyan regions.

The tholeiitic basalts of the Central Atlantic Magmatic province, probably the largest Large Igneous Province so far know on our planet, is instead connected with the beginning of the divergence that will give way to the opening of the Central Atlantic and thus it is more properly linked to the next stage of evolution, the Tethyan stage.

II.2.- Tethys times (Map 6 - Late Norian to Map 13 - Early Aptian)

During this 120 Ma long interval, the Late Palaeozoic oceanic subduction under the thick lithospheric East European Platform goes on, whereas the Pangaea suffers extension between the "Tornquist - Teisseyre line" and a "North African line" (Norian, Sinemurian, Toarcian). In this domain, the extension reaches the oceanic stage (Callovian) and generates a multi-plate structure with a complex kinematics and a sill between the eastern Neo-Tethys and Central Atlantic oceans (the Mediterranean *Seuil*; VRIELYNCK *et al.*, 1996). The Gondwana break-up having being heterochronous, a new Tethys ocean extracts Cimmerian blocks that collide with Asia mainly in Liassic times (see Sinemurian). Both the Atlantic and proto-Indian oceans are synchronous with the complex net of oceanic and thinned crust basins in the Mediterranean *Seuil* (Callovian; Fig. II). The Red Sea and Aden Gulf separate Arabia from Africa in Rupelian times, although during Tethyan times both Arabia and Somalia are frequently transgressed (since Toarcian). The Red Sea and Aden Gulf seem to be located in old lithospheric "lines" which separates domains with different lithospheric characters.

II.2.1.- Palaeoenvironments and main sedimentary environments

The environments are relatively stable during this period and depend mainly upon global factors, both kinematical and biological:

1.- opening of a seaway allowing water transit from Eastern Tethys and Panthalassa to Western Tethys and Panthalassa, too. This grossly speaking east-west seaway has no present day equivalent in the same latitude (DERCOURT *et al.*, 1993);

2.- permanence of latitudinal location (according to the palaeomagnetic accuracy on the maps) e.g., (Fig. III):

- in NW Africa (Agadir) : permanently around 20° N ;
- in Arabia, the northern part of Arabo-Persian Gulf: around 3.5° N (from Norian to Toarcian), around 5° S (from Callovian to Aptian);
- in Central Crimea, around 42° N (from Norian to Toarcian), around 32° N (from Callovian to Cenomanian). Only micro-plates. Only micro-plates involved in orogeny such as the Caucasian and Turkish ones suffering latitudinal movements. Few of them (if any) are plotted on the maps with precise palaeomagnetic data;

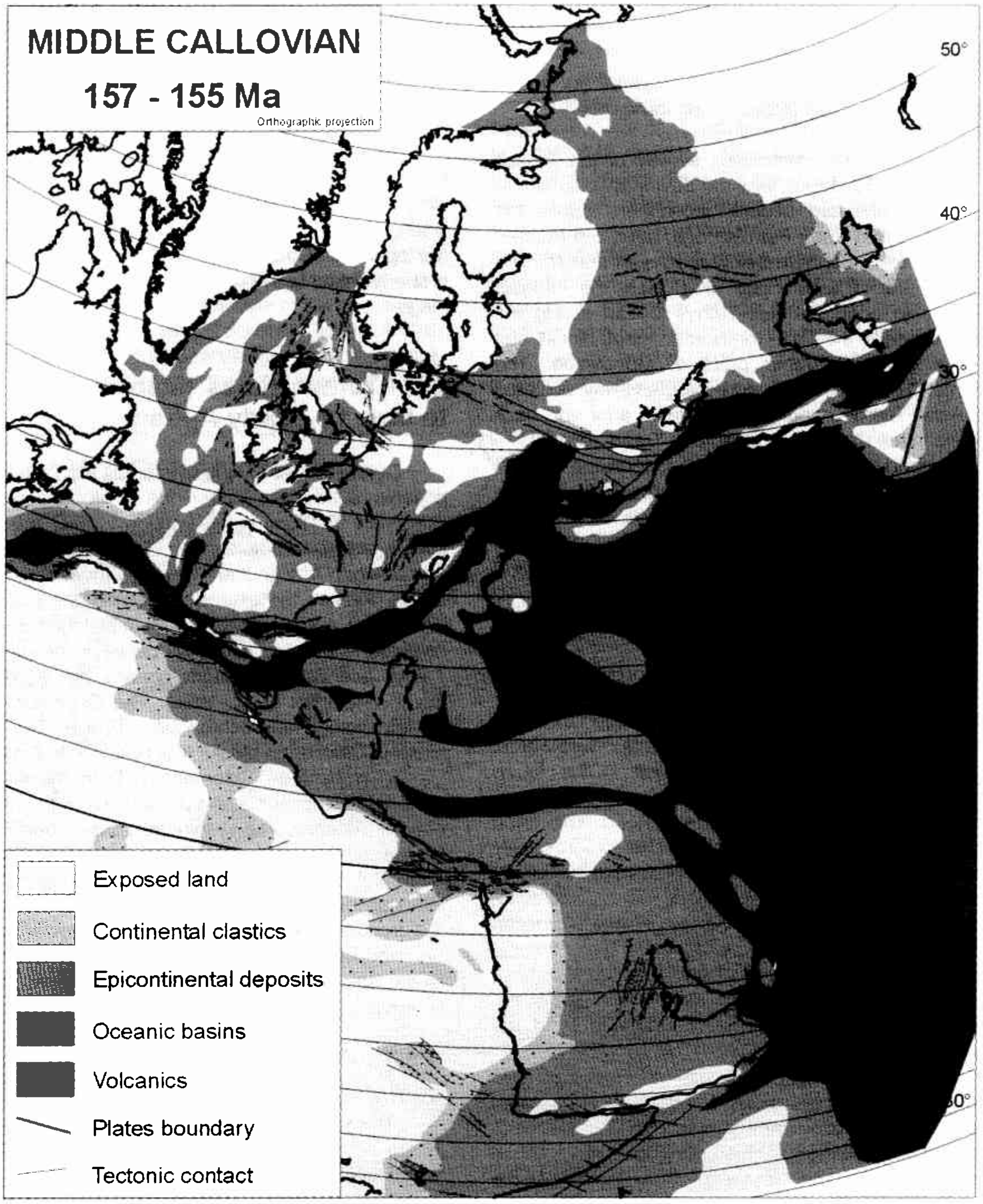

Fig. II: Latitudinal zonation (Middle Callovian).
Note (i) the significant difference between (i) the South Peri-Tethyan platform, Western Europe and the Mediterranean *Seuil,* all included in inter-tropical zone and (ii) North Sea and Eastern European platform.

3.- the changes in palaeoenvironment are dependant mainly on transgression / regression regimes, i.e.:

- the river level equilibrium;
- the importance of water evaporation, very sensitive in this equatorial region (critical for greenhouse effects);
- the rate of spreading and correlative volcanic effects and water chemical composition (hydrothermalism);

4.- the rocks accumulated during Tethyan times are basically bioorganic or with an important organic content and, therefore, dependant on the living species (corals, algae, microplanktonic organisms contained in marine oozes, continental living organisms), biology being an important environmental factor.

Main sedimentary environments

Four domains have specific characters.

1.- The emerged land and marshes

On the emerged land the deeply eroded craton feeds clastic sedimentation and on transitory emergent carbonate platform bauxitogenesis takes place in the suitable latitudes.

- On the northern Peri-Tethyan platform mountain range erosion rules sedimentary activity. Significant rock providers are (i) the Hercynian (Norian to Toarcian), (ii)

Ouralian (Norian to Toarcian) and (iii) Cimmerian orogenies (since Liassic times).

- On the southern Peri-Tethyan platform the African craton tends to be a significant provider, more productive than the northern one, in endoreic, fluviatile and deltaic basins. Age determination is poorly established in continental clastics and the attribution to one map or another is poorly documented as well. Granulometry and volume of clasts are directly related to sea level. Clasts are coarser and abundant in regressive times (Norian - Sinemurian, Tithonian - Hauterivian, Aptian), fine grained and relatively rare during transgressions (Callovian and Cenomanian).

- At the periphery of the continents, at the vicinity of the sea, terrigenous clastic sediments located in equatorial zones (15° N and 15° S latitude) provide fine clasts in shallow environments subject to episodic invasions of seawater and inter-fingered with sediments deposited in fluctuating saline areas.

- When in appropriate areas, bauxites accumulate mainly in karsts. Some alterites are also to be found on recently obducted ophiolites (Aptian). The main phases of bauxitogenesis corresponds to the most important transgression.

2.- Epicontinental deposits

On cratonic basements, under tropical and subtropical climatic zones, coastal deposits, mainly terrigenous, are interbedded with evaporites which concentrate substantially in equatorial climatic zones (15° N or 15° S), e.g., Sinemurian, Toarcian, Kimmeridgian, Tithonian).

As for the main clastic production, carbonate prevails in deposits where fossils species are scarce but individual organisms abundant.

In permanent marine domains, shallow water carbonate platforms are covered by limestone frequently dolomitised. They define supra-tidal to shallow sub-tidal environments. Ramp carbonate platforms develop from shallow to deeper environments. They are widespread at all latitudes. Shelf edges in inter-tropical climatic zones can be fringed by reefal buildups presenting a high diversity of organisms.

Pelagic limestones (nannofossil oozes) are associated with minor siliceous tests (sponges, radiolarians), including siliceous nodules. They are locally iron and manganese enriched, their location being mainly under tectonics control (e.g., *Ammonitico Rosso* in all Jurassic maps). Pelagic marls with rhythmic sedimentation (astronomically controlled?) cover epicontinental basins and talus. They are supposed to accumulate as deep as 1000 m, shallower than most of the black shales. These organic black shales of lacustrine or marine origin concentrated in anoxic to mildly oxic basins and are potential source rock for hydrocarbons (Toarcian, Callovian - Kimmeridgian, Aptian, Cenomanian).

3.- Deep basin (oceanic *p.p.*) deposits

The pelagic marls (see above) may be also deposited in deep basins, even in oceanic basins above the CCD. Most of radiolarites, even if some are sedimented in epicontinental environments, are located below CCD. They are time dependant and the maximum deposits are of Kimmeridgian age.

Note: As a consequence of a generalized subduction, no rock representation was given to most of the oceanic basement surface represented on the maps as deep oceanic basins, except for minor obducted areas.

4.- Active subduction environments

The marginal seas involved in this geodynamic setting have the same sedimentary signals than the passive margins presented above, except for the deposits associated with the subduction itself. Here, the main deposits are turbidites (they are not exclusive of active margins since they exist on distal parts of the talus and on the margins of rapid subsiding intra-cratonic areas) and are always present whenever subduction occurs.

These sequences are frequently siliceous argillite alternating with siliceous calcilutite showing a graded bedding both vertically and horizontally. The deposits include olistoliths and oceanic remnants (ophiolitic blocks); they are rich in volcanoclasts and are inter-fingered with lava flows and intruded by micro-gabbros or micro-dacites. In Tethyan times, they are frequent in the South Caspian, Caucasus and Dobrogea.

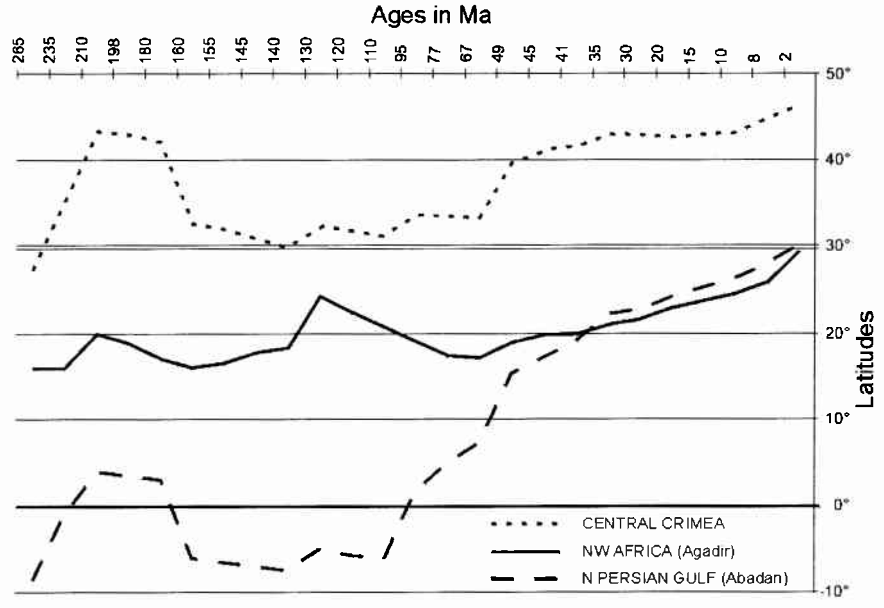

Fig. III: Latitudinal location of Central Crimea, NW Africa and northern Persian Gulf from 265 Ma to Present. Despite plates displacement, each studied area remains in the same climatic zone.

II.2.2.- Structural evolution

The kinematics in Tethyan times is quite simple: extension with predominant crustal thinning between the Tornquist-Teisseyre and North African lines. In the present Atlas, no special attention was given to the future Alpine chain and the Tethys reconstruction (DERCOURT *et al.*, 1993) was modified when necessary, though this was done only in a preliminary form. In the western Tethys, many extensional features were not shown, like the west-facing escarpment of the Central Apennine platform, later reactivated as the Ancona - Anzio Fault. Or in the Balearic area, where no tectonic features are indicated to explain the approaching of Sardinia and Catalonia at closing of the Tethys time. We were also quite sketchy as refers to the Turkish and Hellenic areas, since it was not our direct concern in this Atlas.

Nevertheless, it is worth to mention the behavior of the permanent subducted domain at the south of the East European craton. The Cimmerian blocks collide with the Asian (Turan) platform; an orogen appears (in Aptian times) affected by marginal seas that separate the Cimmerian blocks by thinned or oceanic crust, the main subduction migrating from north to south.

In Tithonian times part of the Neo-Tethys obducted on micro-cratons of the Mediterranean *Seuil* and ophiolites cropped out and flysch soon accumulates (e.g., Bosniac flysch).

Cenomanian (Fig. IV) marks a complete structural change; from Cenomanian on the Alpine times start correlative with the counter-clockwise rotation of the newly established African plate.

II.2.3.- Magmatism

Intracratonic magmatism occurs in association with the crustal thinning in Triassic - Liassic times all over the domains (e.g., West Europe, West Africa, North Sea during Middle Jurassic – not represented in this Atlas).

Representative of the main magmatic activity are the oceanic spreading, the volcanism of island volcanic arcs and the opening of marginal seas associated with subduction in the present-day Caucasian - Dobrogea and Black Sea areas (Sinemurian, Toarcian, Callovian, Kimmeridgian).

II.3.- Alpine times (map 14 - Late Cenomanian to map 24 - Pleistocene)

During this 95 Ma long interval in the southeast, a latest transit-plate (Tauric plate) (RICOU, 1993) migrates from Africa-Arabia to the north in relation with a spreading oceanic ridge (Pamphylian basin) in the west. The northward extension of the Atlantic accretional ridge induces the short-living Biscay ocean between France and Iberia (see Aptian to Lutetian).

Notwithstanding, the *premium motus* in Alpine times is the opening of the South Atlantic Ocean and the individualization of the Africa - Arabia plate with its counter-clockwise displacement. The palaeomagnetic data accuracy evidences, therefore, a strong northward displacement of the eastern part of the plate, and a slight one of the western. This displacement brings about:

- an important accentuation of the pre-existing oceanic subduction under the East European platform during Maastrichtian and, later on, a shift south of the Cimmerian blocks (Iran);

- an initiation of subduction inside the micro-plates constituting the Mediterranean *Seuil*;

- a progressive continent/continent collision generative of mountain ranges (Pyrenees, Alps, Carpathian, Dinarides, Balkan, Pontides, Taurides, Maghrebides, then Zagros) and a general inversion of pre-existing faults inside both Peri-Tethyan platforms, which gives an east-west prominent structure (basins and heights) and inside the Northern platform, instead of the previously diverse directions.

II.3.1.- Palaeoenvironments and main sedimentary environments

The environments change drastically from Tethyan times during which biogenic sedimentation prevailed to Alpine times when clastic deposition took over, with the following results:

1.- the cratons become lesser clastic providers (the break-up of Gondwana allows an easier dispersion of the heat flow underneath Africa - Arabia with the subsequent reduction of the mean altitude and correlatively of the erosion), whereas the incipient orogens provide huge quantities of clasts;

2.- the North Peri-Tethyan platform and east-west basinal alignment separated by highs are the seat of intense sedimentation whose characters depend on:

- marine global stand (transgression vs. regression);

- seaways connecting the main east-west basins and/or boreal or Tethyan oceans;

3.- the progressive closing of the east-west Tethys seaway, which modifies global oceanic current system (dynamic, temperature, geochemical composition, biology...);

4.- a general northward cratonic kinematics, obvious for NE Africa-Arabia, displaces the studied areas across the latitudinal climatic zones e.g., (Figs III, IV and V):

- in Arabia, the northern part of Arabo-Persian Gulf is located at 2° N in Cenomanian, 7° N in Maastrichtian, 19° N in Lutetian, 25° N in Langhian;

- in NW Africa (Agadir), the areas cross from 19° N, in Cenomanian, to 24° N in Langhian;

- in Central Crimea, the sites go from 34° N in Cenomanian to 43° N in Langhian.

Main sedimentary environments

1.-Turbidites and palaeoenvironmental setting

The most typical facies in Alpine times is turbiditic with flysch and molasses. In this work the term “flysch” is used when marine turbidites accumulate in a paleogeographical domain before its main tectonic phase takes place (i.e. the flysch deposits are conformable on the underlying deposits); as for the term “molasses”, it refers to the accumulation of marine, lacustrine or palustrine turbidites in a given palaeogeographic domain after an important tectonic event (i.e. the molasses deposits are unconformable on the underlying deposits) (LORENZ *et al.*, 1993). The flysch basins are elongated

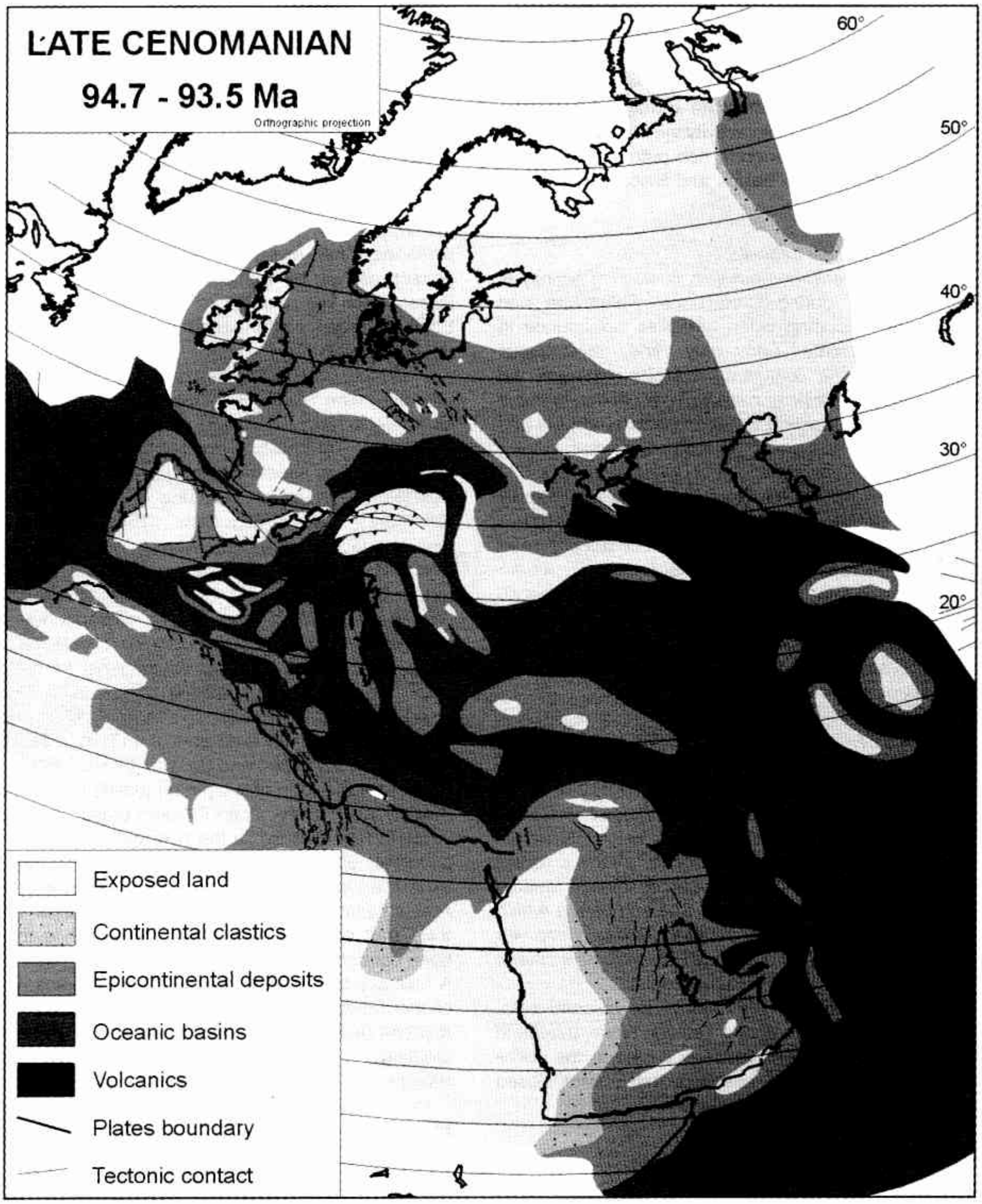

Fig. IV: Latitudinal zonation (Late Cenomanian).

Note (i) the uniform latitudinal zonation of the northern Peri-Tethyan platform vs the southern Peri-Tethyan one, (ii) the tectonism inside the Mediterranean *Seuil*, (iii) the marginal seas related to the Tethyan subduction zone.

(several thousands km long) and relatively narrow (100 km); they migrate in front of the belts preceding them and are mobile. The turbidites are neither characteristic of a particular palaeoenvironment domain, nor of a depth or the nature of the basement, but are rather indicative of the subsidence rate and space availability (e.g., the Late Cretaceous Paleocene helminthite flysch in front of the North Alps micro-plate accumulates first on the Ligurian oceanic crust; then, during the Paleocene, by reduction of space related to subduction, the flysch basin migrates on the continental crust during the Palaeocene in the Briançonnais zone; after that, in Rupelian times in the Dauphiné zone, and as molasses in a foreland basin during Langhian and Tortonian).

Turbiditic flysch and molasses accumulate in marine or brackish waters (fluctuating salinity). These predominant deposits in the east-west alignment of basins are constitutive of the Para-Tethys in the Northern Peri-Tethyan platform and in the seaway connecting the

Para-Tethys complex domain with Boreal, Tethyan or Mediterranean open seas. The subsidence rate is high, abundant lacustrine deposits alternating with evaporites and marine deposits in temperate climatic zones. The heat flow values and tectonic activity vary from medium to low. All these features generate rich petroleum systems (e.g., Vienna, Precaspian basins and Black and Caspian Seas in Rupelian times).

2.- Epicontinental deposits

The shallow environment with fluctuating salinity is mainly carbonate during Cretaceous, Paleocene and Eocene times, occurring when turbidites accumulate in marine environments and only fine argillaceous sediments reach the continental margins, whereas the cratons are minor detritic providers. An exception to these general established facts appears in both the Iberian and French cratons where the Pyrenees range is emerging and molasses accumulate on the bordering cratons.

Shallow marine carbonate deposits are significantly reduced in Alpine times. The northern Peri-Tethys domains cross the Tropic of Cancer and mean water temperature decreases as a consequence of a slower global spreading rate, and, correlatively, the greenhouse effect lessens. Moreover, the opening of seaways connecting Tethys waters with Boreal ones enhances this water temperature decrease. Shallow carbonate deposits concentrate on the African - Arabian shores.

The hemipelagic marine environment is dominated by chalk in the Northern Peri-Tethyan platform and presents a spectacular development in Campanian times. These deposits accumulated under 100 to 600 m water depth and by mass flow can reach very deep basins.

Deep oceanic basin sediments are reduced in Alpine times; most are old ones, some recent (Pamphylian north of Africa in the Cenomanian is filled by sediments, as are the three Black Sea sub-basins which are marginal seas since Cenomanian times). Deep-sea sediments are defined here according to biological criteria not broadly accepted by all palaeobiologists.

Evaporite deposits are found in epicontinental terrigenous domains, like during Tethyan times, though in Late Miocene huge accumulation occurred in the entire Mediterranean basin when both sea gates were closed (Arabian/Iranian and Gibraltar gates).

II.3.2.- Structural evolution

From Cenomanian on, tectogenesis and orogenesis create the Alpine orogens. As concerns the Tethys oceanic domains this has been presented previously (*Atlas Tethys*, Dercourt *et al.*, 1993) and is not within the scope of the present work. In the Peri-Tethys Atlas we have established the consequences of the Tethyan oceanic and margins evolution and subsequent response of the Peri-Tethyan platforms.

The Cenomanian is a key map of the Peri-Tethys Atlas. At the end of Tethyan times, the extension is still active and an incipient compression marks the beginning of Alpine times, accompanied by (i) a prevailing extension in the North Atlantic Ocean, associated with oceanic basins (e.g., Bay of Biscay, Pamphylian and intra-Iranian cratonic basins), as well as in the Africa-Arabia craton (e.g., Tunisia, Egypt, North Arabia, ...); (ii) a compression affecting old oceanic basements of the northernmost micro-plate in the Mediterranean *Seuil* (e.g., Internal Alps) and folding appearing (Western Approaches, Celtic Sea, North Sea basins) in the northern cratonic Peri-Tethys (e.g., Rhenish, Bohemian, Sudetes massifs becoming coalescent by fault inversions).

Since the Maastrichtian, compression dominates (i) by way of the involvement of the Mediterranean *Seuil* micro-plates, while in the north and south Peri-Tethyan platforms fault inversion produces folding (Laramian phase), and (ii) by multiplication of subduction zones in the southern rim of all micro-plates present between the East European platform and the Africa - Arabia plate (Mediterranean *Seuil*).

During Lutetian times, collision continues. All the Mediterranean *Seuil* micro-cratons are thrust northwards on the thinned European craton (south-west of the Teisseyre - Tornquist permanent line) and transtensional incipient basins appear (West European rift). On the thick East European craton, a string of sedimentary basins forms an east-west depression. The permanent oceanic multi-subduction zone develops wide marginal seas (Black Sea, Great Caucasian, Caucasian basins). The former Cimmerian blocks resume collisional activity.

In Rupelian times, as a consequence of the continuing north-south displacement (collision of the Iranian micro-plate and the easternmost Mediterranean *Seuil* micro-plate with the East European platform), the string of east-west basins shaped inside the East European platform incorporates in an unique sedimentary domain the Black Sea and Caspian basins. A new feature appears in the form of a triple rift junction that affects the Africa-Arabia craton. A north-south branch generates an incipient Red Sea, while the northwest-southeast branch generates an incipient Gulf of Aden and a rift system well expressed in the Djibouti Gulf. As a result, a brand new Arabian plate is formed. The three different branches were part of the Tethys times history since they have been controlling the sedimentary domain since Callovian! A new subduction of the oceanic basin begins with parts of the Mediterranean *Seuil* subducting underneath the Iberian-French craton sealed by the Pyrenees, and creating a new oceanic marginal sea, the Algeria - Provence one.

From Burdigalian times on, the West and Central European Alpine mountain chains (southwest of the Teisseyre - Tornquist line) emerge and are eroded, thus feeding fluvio-lacustrine clastic basins; at the northeast of the Teisseyre - Tornquist line, east-west structural elements dominate. The Caucasian range follows the same trend (fold and southward vergence nappes). The Turkish block of Mediterranean *Seuil* origin and the Iranian of Cimmerian origin participate in the Caucasian tectogenesis. The thrust of Iranian Zagros domain on the Arabian micro-plate closes the east-west Tethys seaway, in existence since Callovian times.

During Piacenzian, the same collisional trend continues. The collision between Iberia and Africa closes the western gate of the east-west Tethys Sea. In an isolated form and under open sea influence, the repeatedly desiccation of the Mediterranean leads to Messinian salt accumulation and the resulting diapirism both in deep-sea basins and on their margins.

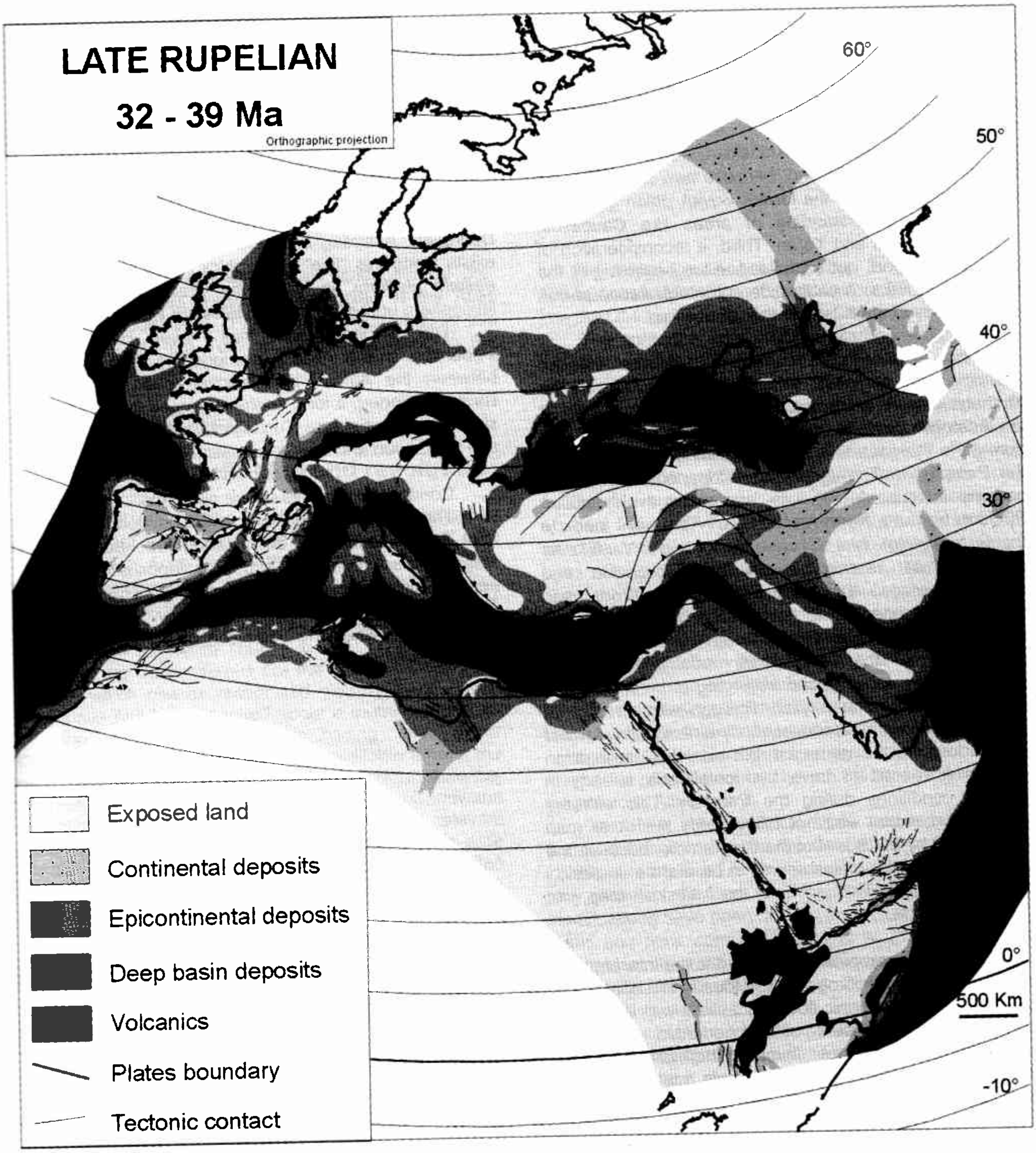

Fig. V: Latitudinal zonation (Late Rupelian).
Note (i) the E-W trending basins in the northern Peri-Tethyan platform, the connections with boreal basin and Tethys remnants, and (ii) the volcanism in Ethiopia - Somalia.

II.3.3.- Magmatism

Intense submarine magmatic activity accompanies the creation of oceanic basins. During Alpine times this activity was moderate. Subduction being a major geodynamic process, andesitic flows are abundant in numerous accreted plates to the East European platform and scarce in the subducting oceanic crust, including the Mediterranean *Seuil* (the present-day Vesuvius or the few volcanoes of the Aegean arc – e.g., Santorini volcano – are samples of this localized activity).

Intra-plate volcanism (mainly alkaline) occurs when rifting systems affect the carbonatic area (e.g., West European rift).

A major volcanic system (hot spot) still highly active nowadays, isolates the Arabian plate from the African one since Rupelian times.

III.- Pending problems

When we started the Peri-Tethys Programme, the basic idea concerning the kinematic was to simply use that of the previous Tethys Project (Dercourt *et al.*, 1993). However, for the Pangaea time this assumption resulted not always viable. First, because the two oldest time slices, Moscovian and Artinskian, were not studied in the Tethys Project. Second, because new field evidence from various states of the former Soviet Union gave new hints to the understanding of areas like Caucasus, Precaspian Basin and Turan. Third, a reconsideration of the Anatolian part, not scheduled at the beginning of the Programme, led to a partly different interpretation of that area, namely of the Kirshehir massif position.

The absence of a conclusive evidence that an ophiolitic belt originally existed to the south of the Kirshehir massif, led to put it on the northern margin of the Anatolian platform, with the Tethys to the north of it during the Pangaea time. The problem is: how large was the Pamphylian Through, when it opened and when it became an oceanic sea-way, separating the Anatolian Platform from the North Arabic Spur? This point leads to the second major point still open. According to Vai & Izart (this volume), responsible for the Moscovian and Artinskian maps, the collision between Africa and Laurussia left open some connections between Panthalassa and Tethys during the Moscovian. Moreover, the propagation of the Neo-Tethys opening during the Artinskian was so advanced according to these authors, that not only the Pamphylian Through was already an oceanic sea-ways and consequently already fairly largely rotated towards the northeast, but also this propagation was so advanced to have the Ionian sea already in oceanic conditions during the Permian. This extreme propagation model was not followed by the other map leaders and by the Board of the Programme, because the oceanisation of the Ionian seems to be more a Jurassic - Cretaceous affair. This is why in the Artinskian map, this area is left blank.

The third open problem is linked to the interpretation of the Caucasus, Dobrogea and Moesia. We considered that the Tethyan oceanic floor, because it was expanding towards NE, underwent oblique oriented subduction under the Eurasian margin, involving therefore also lateral displacements along the active margin and its back arc. This configuration is thought to be important for Caucasus and Turan. This interpretation is partly in contrast with what has been proposed in several models by Nikishin *et al.* (1998a and b, 2000). We moved Moesia to the south-east to have sufficient room to link the Permian sea-way in the middle of Moesia and to accomplish the Middle Triassic palaeolatitudes found by Muttoni *et al.* (2000) for the Pre Balkans. It derives that Dobrogea rocks formed in a fairly large thinned crust basin during the Triassic. With the Cimmerian orogeny, Dobrogea started to rotate towards the East European platform, originating the Norian Flysch. This anti-clockwise rotation allowed the opening during the Jurassic of the Nis-Trojan through to the west. This is also a larger departure from the Tethys Project, which involve a partly different interpretation of the Balkan peninsula. However, its reconsideration was out of the scope of the project and the field work in that area impossible in the present political situation.

The limits of the thin crustal West European - Mediterranean *Seuil* domain.

The northern limit is well established. The Teisseyre - Tornquist line – a coherent set of vertical faults – crosses the North Sea, the Polish Trough, the Carpathian flysch basin (from which the sediments have been ejected and are now the outer Carpathian flysch nappes) and Dobrogea. A strong difference in lithospheric thickness exists along this "line", i.e. a thick lithosphere under the eastern platform as far as the Urals chain vs. a thin lithosphere under Central and Western Europe.

The southern limit has never been directly evidenced. In this Atlas it is suggested as a major limit between the African and Mediterranean lithospheres. Bayonet shaped, a long segment fringes the Maghreb, a short one the Eastern Tunisian coast and another long one the Libyan and Egyptian coasts.

Continental deposits on the platforms.

The coastline delimiting the continental deposits from the epicontinental ones cannot be traced on our maps because these two types of deposits are usually deeply eroded at every regression stage. This is actually a major problem arising at any palaeoclimatic consideration when the extension of the sea acts as a climatic parameter.

1.- MOSCOVIAN (312 - 305 Ma)

G. B. VAI[1] & A. IZART[2]

Unlike all post-Artinskian maps, and because pre-Late Permian times were not considered in the Atlas Tethys (DERCOURT *et al.*, 1993), the two first maps of this Atlas incorporate the Tethyan realm in addition to the Peri-Tethyan platforms.

I.- MAIN FEATURES

I.1.- Time slice definition and resolution

The first map of the Atlas has been conceived to represent the Moscovian stage (or series) of the Late Carboniferous. According to the ICS-IUGS Subcommission on Carboniferous Stratigraphy, the Carboniferous Period is formally subdivided into Early Carboniferous and Late Carboniferous by means of a GSSP defined in Nevada, and ratified in 1996 (Fig. 1.1). The two lower series for the Carboniferous are the Tournaisian and the Visean, which can be recognised globally, although no consensus has been reached about a mutual GSSP. As for the following series subdividing the Upper Carboniferous, no agreement is yet available among the complementary sets derived from Russia, North America, Western Europe, and China, because the Gondwana glaciation during this time interval induced relevant provincialism within the Earth's biota, making world-wide correlation difficult. The Russian Serpukovian to Gzhelian series is tentatively suggested, although no GSSP has been agreed upon yet.

Under such circumstances, we selected a Moscovian time interval on the basis of the most updated and integrated correlation frames defining it in the following biostratigraphic terms. The Moscovian stage (or series) is defined and used for the purpose of this map as roughly represented by the *Aljutovella aljutovica* to *Fusulina cylindrica* fusulinid zones, by the *Idiognathoides marginodosus* to *Idiognathodus obliquus - Neognathodus roundyi* conodont zones, and by the *Paralegoceras* to *Wellerites* ammonoid zones.

About 80% of some 330 validation points used to constrain this map, fulfil the above definition. The remaining 20% are less precisely confined, bracketing the Late Bashkirian to Kasimovian (or even Gzhelian) times. It follows that the peak resolution of many parts of the map corresponds to the stratigraphic amplitude of the Moscovian, but the mean resolution of the map is less sharp and corresponds to Moscovian to latest Carboniferous (or slightly less than the second half of the Late Carboniferous) time.

The numerical time scale for Carboniferous chronological classification down to stage level tentatively adopted for the purposes of this map has a simple pragmatic meaning in order to provide a common frame for the communications of the different contributors. However, as a matter of fact the present state of chronometric calibration of the conventional stratigraphic scale for the relevant time interval is largely unsatisfactory as appears from the very large error bars (from 3 to 9 Ma) affecting individual stage, series and system boundary ages (ODIN, 1994). Especially poor is the knowledge about the Dinantian or Mississippian subsystem. The Late Carboniferous is mainly dated from continental successions not evenly related to the marine ones. A further limitation derives from the low number of formally defined standard chronostratigraphic subdivisions (GSSP) in this time interval. In such a condition, our aim was first e.g., to have a reliable frame of reasonable duration of stages, regardless of precise boundary ages, and secondly to maintain a possible continuity with the time scale used in the previous Atlas Tethys.

Therefore, the numerical scale adopted here (Fig. 1.1) was based mainly on the time scale by ROSS *et al.* (1994), already used in BAUD *et al.* (1993), as a preliminary output of the time scale by MENNING (1995). Some modifications have been derived from CLAOUÉ-LONG *et al.* (1995 and pers. com.), ROBERTS *et al.* (1995) and from attempts at the best fit with HARLAND *et al.* (1990) and ODIN (1994, 1997), GRADSTEIN & OGG (1996) and MENNING (pers. com., 1997, 1999).

As a consequence, the Moscovian map (Fig. 1.2) displays mainly the mean palaeogeographic – palaeotectonic setting from about 312 to about 305 Ma, an interval of about 7 Ma duration, and for some areas an interval of about 15 to 20 Ma.

I.2.- Climatic and cyclostratigraphic trends

The Late Carboniferous to Middle Permian glaciation within Gondwana (WITZKE, 1990; CROWLEY, 1994) had

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1.- Moscovian (321 - 305 Ma)

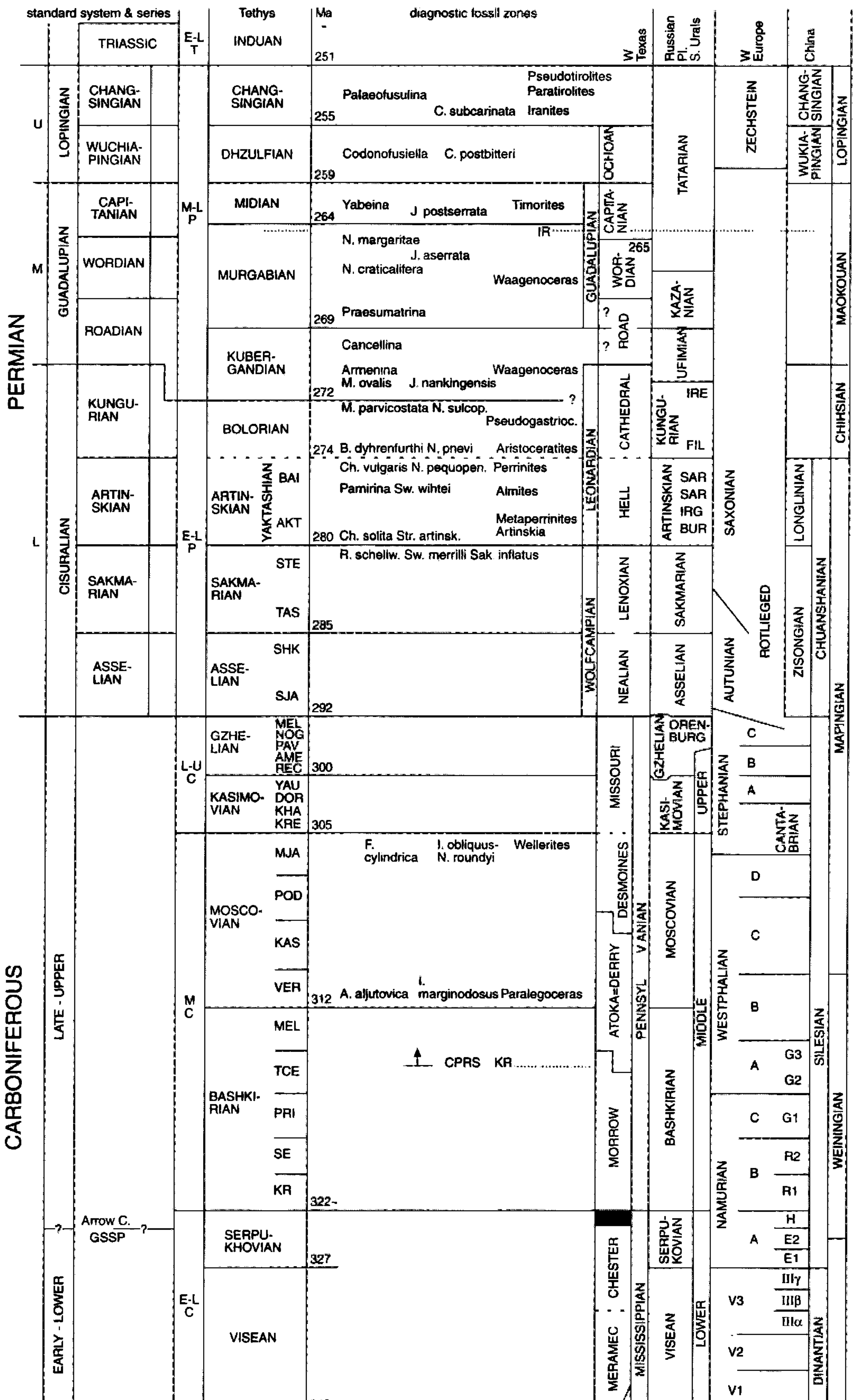

Fig. 1.1: Carboniferous to Permian correlation chart. Finely dotted lines indicate the Kiaman magnetic reversal (KR) and the Illawara reversal (IR); CPRS refers to the Carboniferous - Permian reversed polarity superchrone. The black interval represents the uncertainty in correlating the North American and West European stratigraphic scales through the Arrow Canyon GSSP.

important biotic and cyclostratigraphic consequences. Increasing climatic zonation and related provincialism affected the correlation potential not only among the different provinces (especially between continental and marine realms), but also among different areas inside the same climatic belt (ZIEGLER, 1990; ZIEGLER *et al.*, 1997).Climatic oscillations of different frequency caused fluctuating sea levels and the establishment of cyclical patterns of sedimentation in many large and small shelf areas of the world (Pennsylvanian cyclothems in North America, Russian platform, Carnic Alps, China, Cantabrian Mountains, etc.) (ROSS & ROSS, 1988; MASSARI *et al.*, 1991; VAI & VENTURINI, 1997; IZART *et al.*, 1998; SAMANKASSOU, 1995).

Extensive unconformities punctuating bundles of cycles, and cycles of different frequency provide a tool for correlation among different provinces (IZART *et al.*, 1998).

I.3.- Comments

The world-wide regressive trend characterising the Late Carboniferous as a consequence of the glaciation in the Gondwana is enhanced by the partial accomplishment of the Hercynian orogeny in Europe and parts of the Urals, Mauretanides and Appalachians. In spite of this, the Moscovian represents a peak of relative transgression in the Peri-Tethyan regions, especially concerning the Russian platform, NW Europe, North Africa, as well as North America, South America and the Caribbean region. For this reason, the time interval selected to compile the Late Carboniferous map was the thalassocratic Moscovian epoch, because it provides the largest amount of well dated marine validation points and gives the highest potential of correlation of continental facies by interfingering with marine and transitional ones. The thalassocratic Moscovian regime involved not only the foreland areas of the Hercynian orogen, to both the north (Caledonian peneplane) and the South (Panafrican peneplane), but also the still developing Hercynian belt, especially the foredeep and the chain areas not yet affected by uplift.

I.4.- Environments

A large variety of environments characterises the map, as a consequence of the thalassocratic (high sea-level) regime, and the marked climatic zonation, passing from the glacial belt of South Arabia, Eritrea (between -40° and -50°) and other parts of Gondwana to the everwet belt of Silesia, Ruhr and North America.

Continental mainly erosional to non-depositional environments are extended over a large part of the West Siberia - Kazakhstan continent and the related impressive Uralian - Kazakhstan - Tien Shan volcanic arc; to one half of the Baltic - Fenno-Sarmatian area (including Voronezh and Ukraine highs); one half of the Arabian - Nubian and NW African area, and part of the Hercynian orogenic belt. Reliable evidences of high mountain elevation and relief are found only within the Hercynian belt of Europe (from the Bohemian massif to the French Massif Central and the Iberian axial zone), and especially in the Uralian volcanic arc and the Appalachian internal zone (see for comparison SCOTESE, 1994, fig. 3 with wrong caption in KLEIN & BEAUCHAMP, 1994; ZIEGLER, 1997).

Major continental depositional environments (fluviatile and lacustrine) are practically limited to the northern Gondwana with large circular intracratonic basins in NW Africa and an even larger marginal cratonic basin in East Arabia. The European Hercynian belt is punctuated by numerous small to medium sized, vertical to strike-slip collapse basins. Glacial and fluvio-glacial environments with tillites are found in South Arabia and Eritrea.

Transitional, mainly brackish basins are found at both outer fronts of the Hercynian belt in the British – Danish - Polish foreland basin, the Jerada - Bechar basins in NW Africa, and the Cantabrian basin. Large brackish basins are also associated to the Donetz rift and the Norwegian Sea rift (ZIEGLER, 1989), only partially mapped here.

An evaporitic belt is developed only in the northern part of the mapped area, the large Kara - Tau lagoon (South Kazakhstan), the near Timan archipelago, and the Northern rim of the Baltic shield (between 15° and 35°).

Shallow marine platform carbonates and mudstones largely prevail over silicoclastic coastal deposits. The two major carbonate platforms are represented by the well known Russian platform and the inferred Mediterranean-Turkish platform, up to 2000 km wide and 6-8000 km long, placed North and South of the palaeoequator (a carbonate platform of similar size extended throughout the Amazon - Parnaiba basins in South America outside the map). Carbonate build-ups, bioherms and reefs were limited to or best developed either at the shelf-edge or near to the Carboniferous Palaeo-Tethys, or to the residual oceanic arms of the Uralian belt, the Uralian foredeep and the Precaspian basin.

Deep marine environments are found as discrete foredeeps at the southern Hercynian front, all along the Uralian and Pamir - Tien Shan foredeeps, in the Precaspian basin, and around the Iranian platform. Sediments are mainly represented by thin-bedded turbidites and chert.

Oceanic basins were present in the Carboniferous Palaeo-Tethys and in the not yet consumed segments of the Uralian ocean (NIKISHIN *et al.*, 1996)

II.- STRUCTURAL SETTING AND KINEMATICS

The Moscovian map features a late stage of the Hercynian orogenic cycle. This cycle as a whole is punctuated by major deformation phases (e.g., Acadian - Ligerian, Bretonian, Sudetic, Asturian, Alleghanian, Uralian, encompassing the Mid-Devonian to Late Permian time span). These phases (or group of phases) had different importance in time and different style in space in the various parts of the orogen (ZIEGLER, 1988).

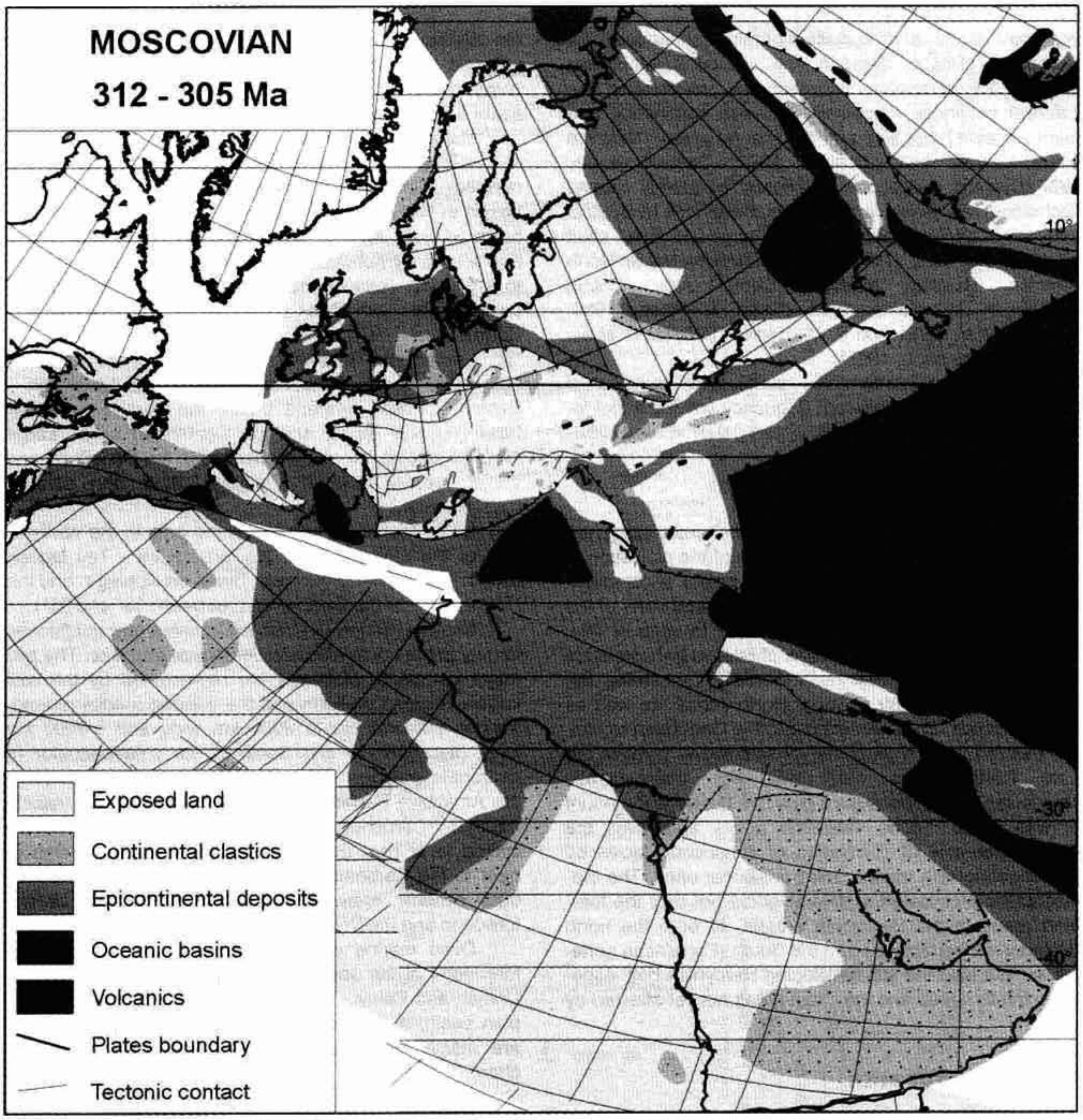

Fig. 1.2: Simplified palaeogeographic map of the Peri-Tethyan area during the Moscovian.

A special feature of the Hercynian orogen is a tripartition in space with two lateral wings, the Urals and the Appalachians - Mauretanides, representing linear orogenic belts, and a central Western European (Variscan) to Mediterranean large body characterised by a dome-like structure (VAI, 1980). The linear belts are interpreted as a result of oceanic subduction and frontal collision; the central body as a result of ensialic block convergence to minor oceanic seaway consumption and microplate/terrane accretion in a large-scale dextral transpressive regime (VAI, 1980; BADHAM, 1982; VAI & COCOZZA, 1986; MATTE, 1986, PIN, 1990). A similar, although transtensional regime was responsible for the progressive post-compressional collapse, granitisation and fragmentation of large part of the Hercynian orogen, especially the central body (ARTHAUD & MATTE, 1977), taking place from Early to Late Permian.

II.1.- Plates and blocks

From the simplified picture given above, it is not easy to list sharply distinct plates and blocks, especially in the mapped area. However, three major plates can be distinguished in lithospheric terms: 1) the oceanic Carboniferous Palaeo-Tethyan plate (part of Panthalassa) and associated microcontinents subducting roughly northward beneath Kazakhstan and the Hercynian Caucasian belt; 2) the West Siberia - Kazakhstan sialic plate overriding; 3) the roughly eastward subducting and

almost consumed Uralian oceanic plate and its East European (Russian platform) passive margin; it was part of the large supercontinent (quasi-Pangaea) originated after consumption of the Rheic (proto-Atlantic) ocean by welding of Gondwana to Laurussia).

However, at the western rim of Palaeo-Tethys it is difficult to assess which ones of the present Middle East basement massifs were welded to the super-continents or were loose microplates or continental fragments.

II.2.- Kinematics and palaeoposition computed

Two alternative scenarios have been used in doing palaeogeographic restorations of the Late Carboniferous to Triassic times. The first one adopts a static approach essentially based on a Bullard-type pre-Jurassic opening Pangaea fit (Pangaea A type). The second one follows a dynamic approach with transition from a Pangaea B to a Pangaea A configuration (MOREL & IRVING, 1981), taking place sometimes during the Carboniferous to Triassic time span.

The palaeolatitudinal control of the blocks reported in the maps during Carboniferous to Permian times is rather poor due to the low number and bad quality of related palaeopole measurements. For this reason the static approach was and is still preferred (SCOTESE, 1984, 1994; SCOTESE & LANGFORD, 1995; ZIEGLER *et al.*, 1997).

However, the dynamic approach is more consistent than the static one with the origin, collapse and fragmentation of the Hercynian orogen as summarised above, implying a nearly continuous dextral strike-slip displacement between Laurussia and Gondwana from Carboniferous to Triassic. The dynamic approach is also more consistent with the set of facies and environmental data used for drawing this and some of the following maps.

Two different mobile solutions have been tested. The first required about 2000 km of longitudinal dextral offset between Gondwana and Laurussia. In the second, an offset of about 800 km was assumed, which is twice the offset already adopted by RICOU (1993) for the Tethys Atlas in the Murgabian map (Late Permian). The second solution fitted the best the available data.

Fold belts, microplates and blocks having a known and reliable Moscovian or Late Carboniferous palaeopole measurement have been positioned accordingly on the map.

The palaeolatitudinal position of Adria was based mainly on the data by MANZONI *et al.* (1989) on the Carnic Alps, showing a reliable (VAN DER VOO, 1993) palaeolatitude of 4°N for the earliest Asselian (following the stratigraphic revision by DAVYDOV & KOZUR, 1997). These data are consistent with the Middle to Late Permian palaeopoles known from the Southern Alps suggesting a continuous northward displacement totalling about 10° from earliest to Mid Permian or 20° from Moscovian.

II.3.- Accuracy and resolution

The palaeolatitudinal resolution of the palinspastic base map is within the confidence of the palaeomagnetic method, which is about 5°. The palaeolongitudinal resolution is undetermined, being the result of assumption and testing in view of the best fitting. The palaeolatitudinal position of the cratons has been adapted to the best fit of the selected kinematics by a displacement less than 1°, that is largely within the resolution of the measured palaeopoles.

Facies and palaeoenvironmental mapping is the result of plotting, interpolating and limiting spot/columnar-like informations provided by some 330 discrete validation points. Each point represents an area with a resolution of about 10 km in diameter. The space density of validation points is rather uneven, with highest frequency in the mountain areas and in oil/mining fields.

In spite of the quite high number of validation points used, the map remains largely inferential, and in part even conceptual, depending on the uneven distribution of validation points, the rather long time interval represented, and uncertainties adopted in the palaeogeographic base map. Major open questions, lack of data, and speculative interpretations are emphasised with blank areas in the map and special mention in the text.

Areas of large uncertainty in the Moscovian map are the Alpine and Alpine - Cimmerian mountain belts and the offshore edges of the present Atlantic passive margins.

The basic palaeolatitudinal position of the cratons in the palinspastic map is dictated by the palaeomagnetic constraints. However, given the poor resolution of this method, any possible attempt to improve the position and to approximate the original size of the mountain belts was made, taking into account the shortening due to post-Moscovian deformations, at least in the range of 100 to 1000 km. A reverse process was experienced in some cases, when available palaeopole measurements made possible an independent check of shortening factors of a belt (as for the Alps and Carpathians).

III.- DEFINITION OF DOMAINS

The following palaeogeographic domains have been distinguished. The more or less elevated erosional highs separating the different basins have not been listed and will be referred to as whenever needed in the next section:

1.- the Late Carboniferous Palaeo-Tethyan ocean;

2.- the West Siberia - Kazakhstan continent and Uralian - Kazakhstan - Tien Shan arc;

3.- the closing arms of the Uralian ocean and Uralian foredeep;

4.- the Precaspian basin (deep marine, turbiditic and shallow platform);

5.- the Russian platform sea (shallow-marine);

6.- the Donetz rift basin (shallow marine to paralic);

7.- the Caucasian - Moesian - Dobrogean - Polish - Oslo basin (shallow marine - paralic to continental);

8.- the syn-tectonic Hercynian foreland basins (paralic);

9.- the intramontane post-tectonic epi-Hercynian basins (continental, collapse to pull-apart);

10.- the Iran - Anatolian basin continuing in the Hellenic - Dinaric - Carnic basin (shallow marine), possibly branching into the Hungarian sea-way with additional connection with the Russian platform sea;

11.- the Apennine basin continuing or branching into the Cantabrian and South Portuguese basins (possibly extending into the South American Amazon basin, outside the map);

12.- the south Peri-Tethyan platform basin (shallow marine, paralic and continental);

13.- the cratonic East Arabian basin (continental);

14.- the intracratonic North African basins (mainly continental).

Domains 6 to 12 (except for 9) are characterised by a *en échelon* pattern following the fragmentation trends of the central part of the Hercynian orogen. They are all open eastward and closed or closing westward. The basins in domain 9, also conforms to the fragmentation directions of the Hercynian orogen, but are much smaller and more confined than the others.

Southwest of the Mediterranean platform basin, the control of the Hercynian fragmentation axes disappears. Circular to elliptical basins of domain 14 show intraplate subsidence accompanied by no or only minor faulting. Others are partly controlled by a different system of faults.

IV.- Description of domains

IV.1.- Late Carboniferous Palaeo-Tethyan ocean

It represented a wide and large western lobe of Panthalassa, left over after the Hercynian suturing of the Uralian and Rheic oceans, and partly confined to the East by the Far East blocks. Clear evidence of northward subduction is found in the Kazakhstan continent (see below) and possibly in the pre-Caucasian Hercynian orogenic belt. A transform margin is assumed at the poorly constrained western edge.

IV.2.- West Siberia - Kazakhstan continent and Uralian - Kazakhstan - Tien Shan arc

The Kuznetsk basin in Russia (Meyen *et al.*, 1996) and the Karaganda basin in Kazakhstan (Litvinovitch *et al.*, 1996) were paralic coal basins from Bashkirian to Gzhelian. The facies were sandstones, coals, lacustrine and marine claystones. The thickness is 1400 m for Upper Carboniferous in Kuznetsk and 500 m in Karaganda. The palaeoenvironments were river, swamp, lake, sea and delta. The flora was of Angaraland-type. The Andean-type subduction is supported by 2 to 4 km-thick accumulation of andesite volcanics in the large Balkash area from Almaty to Karaganda (Nikishin 1998, pers. com.)

IV.3.- Closing arms of the Uralian ocean and Uralian foredeep

The Uralian ocean was almost consumed except for the North Urals and Novaya Zemlya (Nikishin *et al.*, 1996). The long belt of thick andesitic volcanics that punctuates the Urals orogen is evidence of eastward subduction. The northward propagation of the oblique collisional belt and the load of the migrating nappes started the subsidence in the Uralian foredeep.

The southern Urals basin in Russia (Belaya; Proust *et al.*, 1998) was mainly part of this foredeep during Bashkirian to Early Permian. The facies were black shale and turbidites. The thickness was less than 100 m from Moscovian to Gzhelian (Fig. 1.1), suggesting starved basin conditions. The palaeoenvironments were basin and slope.

IV.4.- Precaspian basin (deep marine, turbiditic and shallow platform)

In the Kazakhstan part (Ensepbaev *et al.*, 1998), the border of the Precaspian basin presented from Moscovian to Gzhelian a shallow platform in Janajol at East and Tengiz at South and deep marine basin in Akjar and towards the centre of the basin. The facies in Janajol were limestones, sandstones and claystones and in Akjar black shale and calcturbidites. The thickness exceeding 1100 m in Janajol is less than 100 m in Akjar condensed sequence (starved basin). The palaeoenvironments were a shallow platform in Janajol and a deep marine basin in Akjar.

After the post-Baikalian Early Palaeozoic extensional plate reorganisation (Nikishin *et al.*, 1996), the Precaspian basin formed as a large (over 700 km-wide) near circular cratonic depression. The pulsating Palaeozoic subsidence (rift phases), including a Late Carboniferous to Permian interval, is variously interpreted. Sublithospheric mantle tear-off at the crossing of the eastward subducting European plate, and the northward subducting Palaeo-Tethyan plate is a mechanism more appropriate than or additional to the two-sided thrust-loading (from the South Urals and the Karpinsky swell) as claimed by Nikishin *et al.* (1996). In fact, the structural map of the Precaspian basement (Volozh *et al.*, 1997) shows a geometry poorly affected by the two very heterogeneous fold belts.

IV.5.- Russian platform sea (shallow marine)

Moscovian to Gzhelian deposits were observed in Gubakha near the border of the central Urals basin. The facies are bedded shelf-limestones and build-ups framed by algae and *Palaeoaplysina*. The thickness is near 600 m. The palaeoenvironments were inner, mid and outer ramp with small build-ups.

Moscovian to Gzhelian deposits exist also in the western part of the Russian platform near Moscow (Briand *et al.*, 1998) (Fig. 1.1). The facies are shelf-sandstones, bedded limestones and claystones, presenting a thickness of 300 m. The palaeoenvironments were inner, mid and outer ramp.

Tectonics: stable to slightly extensional structural platform area.

IV.6.- Donetz rift basin (shallow marine to paralic)

It is a survivor of the Mid to Late Devonian Pripyat - Dniepr - Donetz - Donbass - Karpinsky rift (aulacogen).

Sections were chosen near Donetz town in an intermediate area between continental and marine deposits from Moscovian to Gzhelian (IZART *et al.*, 1996, 1998). The facies near Donetz are alternations of sandstone, coal, limestone, marine - lagoonal - lacustrine - claystone, siltstone. The thickness reaches 4826 m near Donetz town, thought it decreased towards the north-west (more continental facies) and increases towards the south-east in the Russian part of Donetz basin (more marine). The palaeoenvironments were river, swamp, inner, mid and outer ramp, marine-lagoonal and lacustrine deltas near Donetz town. Tectonics: rift-related extension, possibly connected with the north-dipping Palaeo-Tethys subduction (ZIEGLER, 1988).

IV.7.- Caucasian - Moesian - Dobrogean - Polish - Oslo basinal belt (shallow marine - paralic to continental)

The North Caucasus basin (CHERNYAVSKY *et al.*, 1975) and the Zonguldak basin (North Turkey, KEREY *et al.*, 1985) were limnic coal basins from Westphalian to Stephanian (Fig. 1.1). The facies is formed of alternations of conglomerate, sandstone, coal, claystone and siltstone with a thickness of 860 m for Caucasus and 475 m for Zonguldak. The palaeoenvironments were river, swamp, lake and delta. Moesia shows Westphalian fluvial, limnic, deltaic coal-rich, and shallow-marine carbonate palaeo-environments. An alluvial plain and a narrow silicoclastic marine band occupied Dobrogea that was separated from Moesia by an emergent Hercynian front (YANEV *et al.*, comm. pers. 1998; SEGHEDI & PANA, comm. pers. 1998). This means that at least the western part of Moesia was affected by the Hercynian orogeny (VAI, 1980).

The Lublin basin (ZDANOWSKY & ZAKOVA, 1995) in Poland was a paralic coal basin during the Westphalian A and B and limnic coal basin during Westphalian C and D. The facies are alternations of sandstone, coal, claystone and siltstone. The thickness is 1012 m. The palaeoenvironments were river, swamp, lake and delta. Tectonics: basin located near the Tornquist normal fault.

The Oslo basin (OLAUSSEN *et al.*, 1994) existed during the Moscovian. The facies are conglomerates, sandstones, claystones and limestones. The thickness is 50 m. The palaeoenvironment was a marine platform and deltas. Tectonics: extension, pre-rift phase.

Fully marine silicoclastic to carbonate conditions were related to cyclic seal-level fluctuations during the Permo-Carboniferous glaciation in Gondwana. Nevertheless, three possible marine connections for this narrow sea-belt are suggested: with the Palaeo-Tethys to the south-east, with the Russian shelf through the Donetz rift to the north-east and with the Greenland - Norwegian sea to the north-west. The EW-trending eastern part of the belt was located inside and parallel to the North front of the Hercynian orogen separating Moesia from North Dobrogea (thrust-top basin). The NW-SE-trending western part of the belt was just outside the Hercynian front running parallel to the Tornquist extensional belt.

IV.8.- Syn-tectonic Hercynian foreland basins (paralic)

A large paralic coal-basin covered all north-west Europe during Westphalian in England (RAMSBOTTOM *et al.*, 1978), Wales (ARCHER, 1968), North France (BOUROZ *et al.*, 1964), Belgium (PAPROTH *et al.*, 1983), Netherlands (GELUK, 1997), Germany (Ruhr; FIEBIG, 1969) and Poland (Silesia; ZDANOWSKY & ZAKOVA, 1995). The facies were alternations of sandstone, coal, marine - lagoonal - lacustrine claystone and siltstone. The thickness was 500 m in England, 1600 m in Wales, 1280 m in North France, 1350 m in Ruhr, 1392 m in Netherlands and 1074 m in Silesia. The palaeoenvironments were river, swamp, sea, lagoon, lake and deltas. In these basins, the last marine bands occur in the lower part of Westphalian C (Fig. 1.1). This basin filled an elongated foredeep to foreland flexural depression at the north front of the Hercynian chain.

IV.9.- Intramontane post-tectonic Hercynian basins (continental, collapse to pull-apart)

Limnic coal basins existed in Lorraine (DONSIMONI, 1981), Saar (KORSCH & SCHÄFER, 1995), the Czech Republic (OPLUSTIL & PESEK, 1998) and Slovakia (VOZAROVA, 1998) during the Westphalian inside the Hercynian chain in Saxo-Thuringian and Moldanubian zones. The facies are alternations of conglomerate, sandstone, coal, claystone and siltstone. The thickness is 3100 m in Lorraine, up to 300 m in Czech Republic and Slovakia for the Westphalian C and D. The palaeo-environments were river, swamp, lake and delta.

Limnic coal basins existed in Saint Étienne (DOUBINGER *et al.*, 1995), Lorraine (DONSIMONI, 1981), Saar (KORSCH & SCHÄFER, 1995), Saale (SCHNEIDER, 1996), the Czech Republic (OPLUSTIL & PESEK, 1998) and Slovakia (VOZAROVA, 1998) during Stephanian in Saxo-Thuringian and Moldanubian zones. The facies are alternations of conglomerate, sandstone, coal, claystone and siltstone. The thickness is 5600 m in Saint-Étienne, 1000 m in Lorraine, 2340 m in Saale, up to 500 m in the Czech Republic and Slovakia for the Stephanian. The palaeoenvironments were river, swamp, lake and delta. Extensional collapse in the internal zones of Hercynian Europe prevailed over the beginning of the transtension.

IV.10.- Iran - Anatolian basin continuing in the Hellenic - Dinaric - Carnic basin (shallow marine), possibly branching into the Hungarian seaway with

additional connection with the Russian platform sea

The Elbourz basin in Iran (JENNY *et al.*, 1978), the Anatolian - Hellenic - Dinaric basins (DEMIRTASLI, 1990; PAPANIKOLAU & SIDERIS, 1990; RAMOVS, 1990), as well as the Carnic Alps basin (VAI & VENTURINI, 1997; KRAINER & DAVYDOV, 1998), and the Transdanubian - Bükk basin presented platform deposits from Moscovian to Gzhelian (Fig. 1.1). The facies are alternations of limestone and sandstone in Elbourz and conglomerate, sandstone, claystone and limestone in Carnic Alps and other areas. Faunal and floral relations were very strong with the Russian platform and the Tethyan regions (VAI, 1994, 1997). This is consistent with the restored position of the branching Hungarian seaway facilitating a direct connection with the southern Urals and the Russian platform. The thickness is 162 m for Elbourz and 1542 m for Carnic Alps. The palaeoenvironment was a marine platform. Hercynian deformation was absent in Iran, and gradually increased from Anatolia to the Dinarides and the Carnic Alps where a marine molasse stage is well developed. The Hellenic and South Turkey basins (Jadar, Vardar, Hydra, Lesvos, Chios, Kavaklidere, Kisanta, Bitlis) developed near the South front of the Hercynian chain or in its foredeep, and bear evidence of strike-slip faulting especially in the Carnic Alps (VAI & VENTURINI, 1997).

IV.11.- Apennine basin continuing or branching into the Cantabrian and South-Portuguese basins (possibly extending into the South American Amazon basin, out of the map)

Tuscany and Southern Apennine (Gargano) basins (PASINI & VAI, 1997) presented platform and deep marine deposits during the Moscovian. The Cantabrian basins exhibited platforms in Picos de Europa (VILLA, 1985) during Moscovian and Kasimovian, paralic deposits in central Asturian coal basin (DIAZ, 1983) during Westphalian, and paralic - limnic coal basin in NW Spain basins (BOUROZ *et al.*, 1972 ; WAGNER & WINKLER-PRINS, 1985) during Stephanian. The South Portuguese basin (OLIVEIRA *et al.*, 1983) was deep marine up to Early Westphalian. The facies are limestones and claystones in part of Tuscany and Picos de Europa, alternations of sandstone, coal, marine - lagoonal - lacustrine claystone and siltstone in coal basins, turbidites in part of Tuscany and the South Portuguese zone. Faunal relations were very close to that of the South American Amazon basin (VAI, 1994, 1997). This is consistent with a NW Africa offshore to Florida marine connection (outside of the map) or with possible Bashkirian to Moscovian connection through central West Africa. The thickness is 200 m in the Tuscany platform, 500 to 1000 m in the Tuscany basin, 2546 m in Picos de Europa, 5000 m in Central Asturia, 1025 m in North-west Spain, unknown in the South Portuguese zone. The palaeoenvironments were marine platform in Picos de Europa, part of Tuscany and Southern Apennines; river, swamp, sea - lagoon - lake, delta in paralic-limnic basins; deep-basin in part of Tuscany and the South Portuguese zone. This domain extended over the foredeep basins at the southern Hercynian front, with evidence of strike-slip faults in Cantabria.

IV.12.- South Peri-Tethyan platform basins (shallow marine, paralic and continental)

The Moroccan coal basins were limnic (Sidi Kassem) during Westphalian D or paralic (Jerada) during Westphalian C (DESTEUCQ *et al.*, 1988). The Algerian coal basin of Kenadza (DELEAU, 1951) was paralic during Westphalian C and continental during Westphalian D (Fig. 1.1). The Algerian basin of Mezarif (NEDJARI, 1982) was marine during the lower part of Moscovian and continental during the upper part. Data are missing concerning the presence of Carboniferous under the Mesozoic and Cainozoic nappes and thrust belts of the Rif, Tell and Atlas (NW Africa). The Tunisian basin (Djeffara, Kirchaou; LYS, 1988) was marine during Moscovian and Kasimovian and emerged during Gzhelian. The Libyan basins in Ghadames (COQUEL *et al.*, 1988; MASSA & VACHARD, 1979) and Cyrenaic (VACHARD *et al.*, 1993) were marine during Moscovian and continental or marine during Kasimovian and Gzhelian. The North-eastern Egypt basin (KORA, 1998) was marine during Moscovian, Kasimovian and Gzhelian. The Karadag and Bademli basins in Southern Turkey (MONOD, 1977; DEMIRTASLI, 1990) and Syrian (Sawanet) basins (AL YOUSSEF & AYED, 1992) were marine from Moscovian to Gzhelian. The facies are alternations of conglomerate, sandstone, coal, limestone, marine - lagoonal - lacustrine claystone, siltstone in the Moroccan and Algerian coal basins. The facies are limestones and marine claystones in Tunisia and Libya, limestones and sandstones in South Turkey and marine claystones and sandstones in Egypt (Wadi Araba and Sinai) and Syria. The thickness is 425 m in Djerada, 1670 m in Kenadza, 280 m in Tunisia, 180 m in Ghadames and 313 m in Cyrenaic (Libya), 176 m in Egypt, 102 m in Southern Turkey and 1000 m in Syria. The palaeoenvironments were river, swamp, sea - lagoon - lake and deltas in paralic and limnic coal-basins, continental and marine platform in Tunisia, Libya, Egypt, South Turkey and Syria. Most of these are stable platform basins developing in the southern Panafrican foreland of the Hercynian orogen. The North Syria basin only reflects the beginning of the Palmyra rift.

IV.13.- Cratonic East Arabian basin (continental)

During Late Westphalian and the Stephanian, Saudi Arabia (OWENS & TURNER, 1995; AL LABOUN, 1993) and Oman (LOVE, 1994) presented continental deposits unconformably overlying Precambrian to Devonian rocks (evidence of intraplate compressive stress conditions in the southern foreland of the Hercynian belt). The facies are conglomerate, sandstone, claystone and diamictites. The thickness is 125 m in Saudi Arabia and 92 m in Oman. The palaeoenvironments were continental with tillite and alluvial-plain deposits. Facies distribution (SCHANDELMEIER *et al.*, 1997) suggests that the basin

extended over most of the Arabian peninsula, and its present western limit would relate mainly to Neogene updoming and erosion along the Red Sea. The Arabian platform was stable or slightly uplifting, except for its North edge along the Sinai and Palmyra rifts. Transpression is reported at the SE edge of Oman (SCHANDELMEIER *et al.*, 1997).

IV.14.- Intracratonic North African basins (mainly continental)

Continental deposits are exclusive of the large NW African Tindouf, Reggan (CONRAD, 1985), Taoudenni and Iullemedden (LEGRAND-BLAIN, 1985) basins. The NE African Murzuk - Djado and Kufra - Erdi basins (MASSA, 1985; LEGRAND-BLAIN, 1985) were still largely marine or passing from marine to continental conditions due to Late Carboniferous eustatic sea-level oscillations. Facies are mainly shaly to sandy and sometimes carbonate (limnic to the west and marine to the east). The thickness is 100 to 200 m for the Taoudenni, 500 to 700 m for the Tindouf, 80 m for the Reggan, about 200 m for the Murzuk - Djado, and 200 to 300 m (the "glacial formation") for the Kufra basin.

Tectonic control by NS-trending normal faults (present grid) is clear for the Iullemedden and possible for the Murzuk - Djado and Kufra - Erdi basins. It is tempting to assume a Moscovian shallow marine connection of the south Peri-Tethyan basins with the South American Amazon and Parnaiba basins (outside the map) via the Murzuk - Djado basin. This would be consistent with the South American affinity of the Tuscan corals and conodonts (VAI, 1997; see above).

2.- ARTINSKIAN (280 - 273 Ma)

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Because pre-Late Permian times were not considered in the Atlas Tethys (DERCOURT *et al.*, 1993) the Artinskian map of this Atlas incorporates, as it does for the Moscovian, the Tethyan realm in addition to the Peri-Tethyan platforms.

I.- MAIN FEATURES

I.1.-Time-slice definition and resolution

Two maps of the Atlas concern the Permian period, Artinskian and Wordian. The Permian stratigraphic subdivisions have been undergoing a significant upgrading in the last few years. Traditionally the Permian was subdivided into two series. The lower part was largely based on the Pre-Uralian basin succession and the sequence, bottom to top, formed by the Asselian, Sakmarian and Artinskian stages, was largely followed, whilst stages based on North America and China had only local meaning.

The progressive transition from marine to marginal and even continental environments in the Perm - Kazan area of the Urals, and consequently the difficulties in their correlation, led to identify four different regional scales (Uralian, Tethyan with stages defined in Transcaucasia and Pamir, North American and Chinese). The increasing provincialism of the biostratigraphic features introduced a number of bias in the interregional correlations and resulted in the stratigraphic nomenclature of the Permian becoming more and more confused. Only the lower three stages obtained a fairly large consensus and use.

The second map of the Atlas (Fig. 2.1) represents the Artinskian stage of the Early Permian series. In fact, according to the ICS-IUGS Subcommission on Permian Stratigraphy, the Permian Period/System is formally subdivided into three series (JIN *et al.*, 1997). This subdivision was also adopted by the COMMISSION OF THE GEOLOGICAL MAP OF THE WORLD (1998). The Lower Series, named Cisuralian is based on the Ural succession, the Middle Series named Guadalupian on the North American successions and the Upper Series, named Lopingian, on the Chinese successions. With such a diplomatic partition, the only excluded were the Tethyan successions, which perhaps are the best, but are cropping out in remote and politically unstable areas after the wreckage of the Soviet Union. Figure 1.1 (Moscovian map) illustrates the present status of correlation for the Permian.

We selected an Artinskian time interval on the basis of the most updated and integrated correlation frames defining it in the following biostratigraphic terms. The Artinskian stage *sensu lato* is used for the purpose of this map as comprising the Artinskian stage *sensu stricto* and the Bolorian stage, roughly corresponding to the *Chalaroschwagerina solita, Pamirina, Ch. (="Pseudofusulina") vulgaris, Misellina dyhrenfurthi* and *Misellina parvicostata* fusulinid zones, to the *Streptognathodus artinskiensis, Sweetognathus whitei, Neostreptognathodus pequopensis* and *N. leonovae* conodont zones. It is characterised by the new fusulinid genera *Mesoschubertella, Toriyamaia, Praeskinnerella, Darvasella,* and by the ammonoid genera *Artinskia, Kargalites, Almites, Cardiella, Paragathiceras, Aristoceratoides, Pseudogastrioceras.* The selected late Early Permian time has some advantages because its duration is almost similar to that of the Moscovian. As a matter of fact, it is closely equivalent to the Chinese Chihsian stage and practically overlaps with the Yachtash - Bolor faunal step of LEVEN (1993), comprised between the top of Sakmarian and the base of Kubergandian. Moreover, it minimises the imprecision deriving from the poor correlation and definition level of the individual stages involved.

Around 80% of some 270 validation points used to constrain this map, fulfil the above definition. The remaining 20% are less precisely confined, bracketing the Asselian to Artinskian times. It follows that the peak resolution of many parts of this map is the stratigraphic amplitude of the Artinskian, but the mean resolution of the map is less sharp, including the earliest Permian up to the Artinskian.

The numerical time scale for the Permian chronological classification down to stage level tentatively adopted for the purposes of this map has a simple pragmatic meaning, which is to enable a common frame for communication among different contributors. However, as a matter of fact the present state of chronometric calibration of the conventional stratigraphic scale in the time interval concerned is largely unsatisfactory as appears from the very large error bars (from 5 to 8 Ma) affecting individual

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stage, series and system boundary ages (ODIN, 1994). A further limitation derives from the still poor level of formally defined standard chronostratigraphic subdivisions (GSSP) at the stage level in this time interval. In such a condition, our aim was firstly to have a reliable frame of reasonable duration of stages, regardless of precise boundary ages, and secondly to maintain a possible continuity with the time scale used in the previous Tethys Program.

Therefore, the numerical scale adopted here (Fig. 1.1) was based mainly on the time scale by ROSS *et al.*, (1994), already used in BAUD *et al.* (1993), as a preliminary output of the time scale by MENNING (1995). Some modifications have been derived from CLAOUÉ-LONG *et al.* (1995 and pers. com.), ROBERTS *et al.* (1995) and from attempts at the best fit with HARLAND *et al.* (1990), ODIN (1994), GRADSTEIN & OGG (1996) and MENNING (pers. com., 1997, 1999).

Consequently, the Artinskian map displays mainly the mean palaeogeographic - palaeotectonic setting from about 280 to about 273 Ma, an interval of about 7 Ma duration, and for some areas an interval of about 20 to 23 Ma, down to the base of the Early Permian (about 296 Ma).

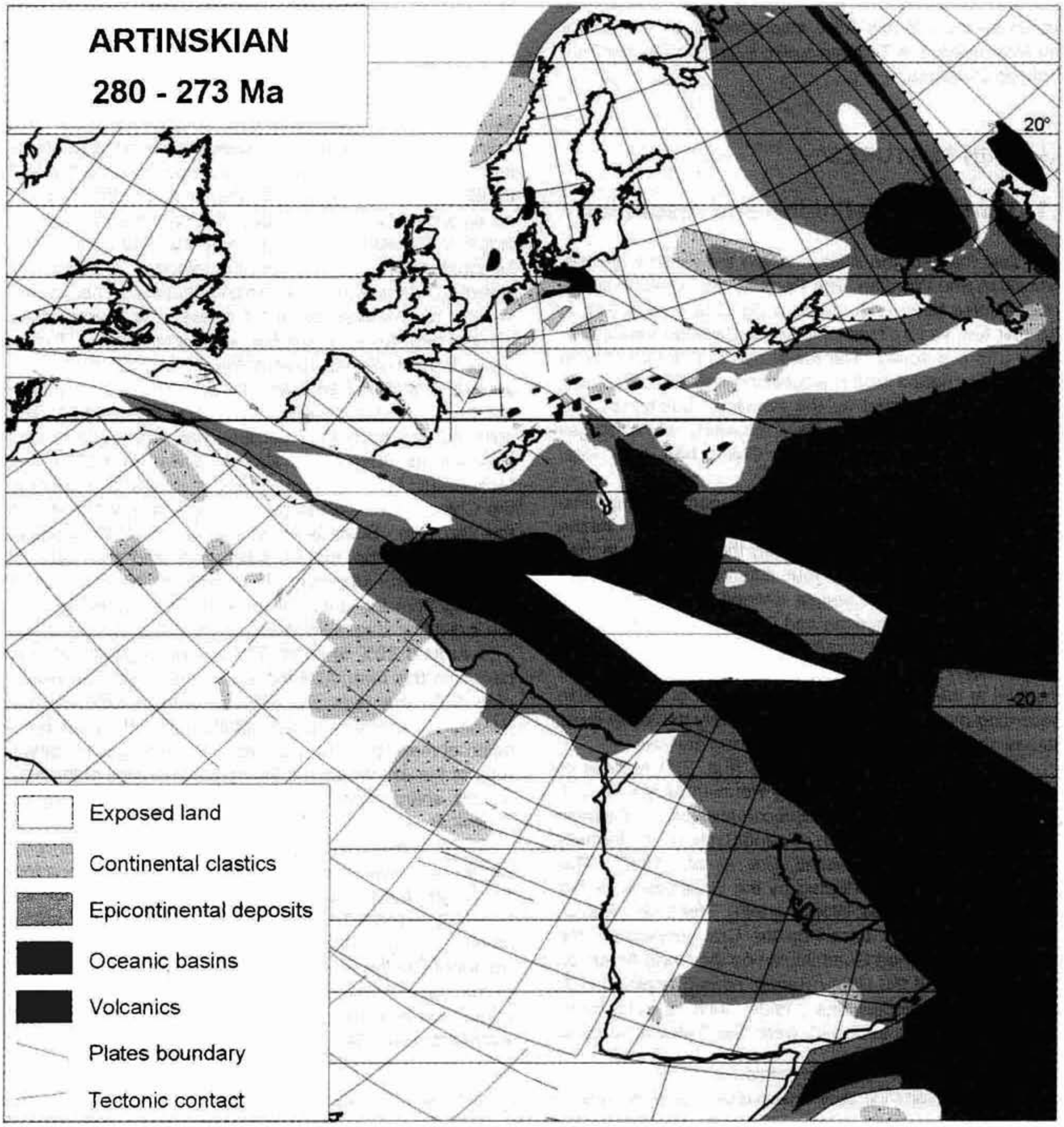

Fig. 2.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Artinskian.

II.- STRUCTURAL SETTING AND KINEMATICS

The Artinskian map is dominated by the Early Permian dextral megashear (ARTHAUD & MATTE, 1977), transecting and fragmenting the entire central body of the Hercynian orogen (see Moscovian expl. note). This megashear culminated with the Oman - Levantine - Sicilian-Texan dextral transform rift. Its eastern part developed in oceanic to sub-oceanic conditions, while the western part split the sialic Mauretanides - Appalachian belt along its axis, interrupting the convergence for some tens of Ma. Compression resumed during Late Permian concurrently with the initiation of the Somaliland rift (see above), and the orogen was completed with the Permian. Evidence of this rift (outside the map) is found along the entire Malagasy marine basin during Early to Mid Permian, as well as in the possible marine connection of the Congo basin with the South American Paranaibo and Amazon basins (WISSER, 1997; WOPFNER & CASSHYAP, 1997).

Along the Urals belt, compression was still going on. Northward subduction of the oceanic Palaeo-Tethys was proceeding, especially beneath the Pamir - Tien Shan arc, as shown by the thick Balkash Lake calc-alkaline volcanics.

II.1.- Plates and blocks

From a kinematic point of view, the setting has changed during the transition from the Moscovian to the Artinskian. The former quasi-Pangaea supercontinent was split into two parts displacing dextrally each other along the Oman - Levantine - Sicily - Texan rift. Three major plates can be recognised in lithospheric terms: 1) the oceanic Permian Palaeo-Tethyan plate (a part of Panthalassa) subducting roughly northward beneath Kazakhstan; 2) the Laurasian continental plate displacing dextrally in respect to the 3) Gondwana continental to oceanic plate. At its NE edge, the Turkish and Iranian blocks (together with other ones to the east, see Atlas Tethys; DERCOURT *et al.*, 1993) had already started their anticlockwise displacement as Gondwanan fragments. Similarly, the Indian plate also started rifting off Gondwana.

II.2.- Kinematics and palaeoposition computed

Following the dynamic approach discussed for the Moscovian map, the palinspastic base map of Artinskian was drawn assuming a dextral displacement of about 300 km.

Fold belts, microplates, and blocks having reliable Artinskian or Early Permian palaeopole measurements (VAN DER VOO, 1993) have been positioned accordingly on the map. For the palaeolatitudinal position of Adria see the explanatory note of the Moscovian.

II.3.- Accuracy and resolution

For all general remarks, the reader can refer to the explanatory note of the Moscovian map. Areas of large uncertainty are the Alpine - Cimmerian mountain belts plus Moesia, and the offshore edges of the present Atlantic passive margins.

III.- DEFINITION OF DOMAINS

The following palaeogeographical domains have been listed. They largely overlap, thought not completely, those of the Moscovian map:

1.- the Early Permian Palaeo-Tethyan ocean;

2.- the West Siberia - Kazakhstan continent and the Uralian - Kazakhstan - Tien Shan arc;

3.- the Uralian chain and foredeep;

4.- the Precaspian basin (deep marine, turbiditic);

5.- the Russian (or Moscow or East European) platform sea (shallow marine to evaporitic);

6.- the Donetz rift basin (evaporitic to continental);

7.- the Caucasian - Moesian basin (shallow marine to continental);

8.- the Oslo basin (rift);

9.- the intramontane post-tectonic epi-Hercynian basins (continental, collapse to pull-apart);

10.- the Iranian - Anatolian basin continuing in the Hellenic - Dinaric - Carnic basin (shallow marine);

11.- the Apennine basin;

12.- partly inferred, the Oman - Levantine - Sicily oceanic sea-way (Permian Tethys), continuing into the Sicani - Djeffara deep-marine basin and possibly extending to a sialic sea-way connecting the western Tethys with the Caribbean and Panthalassa areas;

13.- the south Peri-Tethyan platform basins (continental, shallow to deep marine);

14.- the cratonic East Arabian basin (continental, shallow marine to evaporitic);

15.- the intracratonic North African basins (mainly continental).

Domains 6 to 12 (except for 9) are still characterised by a *en échelon* pattern following the fragmentation trends of the central part of the Hercynian orogen. They are all open eastward and closed or closing westward. The basins in domain 9 clearly reflect a conjugate pattern of fragmentation overlapping the Moscovian one, but developing independently from the Hercynian orogenic collapse, as shown by the distribution of the widespread bimodal volcanism both inside and outside the Hercynian orogen, as a consequence of the Early Permian rift. The axis of the oblique rift is located along domain 12. Minor rift-induced fragmentation effects in the Gondwana plate are present in the SE part of the Kufra - Erdi basin only. Circular to elliptical basins of domain 15 are still subsiding, with decreasing control by the ancestral faults having the present meridian trend.

IV.- DESCRIPTION OF DOMAINS

IV.1.- Early Permian Palaeo-Tethyan ocean

It continued its northward subduction beneath the Kazakhstan continent, and narrowed in comparison to the Moscovian.

IV.2.- West Siberia - Kazakhstan continent and Uralian - Kazakhstan - Tien Shan arc

The Kuznetsk basin in Russia (MEYEN *et al.*, 1996) was a paralic coal basin during Asselian. The facies were sandstones, coals, lacustrine and marine claystones. The thickness was 300 m. The palaeoenvironments were river, swamp, lake, sea and delta. The Andean-type subduction of the Palaeo-Tethys ocean beneath the Kazakhstan - Tien Shan arc continued with the accumulation of thick andesitic volcanics in the Balkash Lake area (see Moscovian).

IV.3.- Uralian chain and foredeep

During the Early Permian, the South Ural basin in Kazakhstan (Aidaralash; DAVYDOV *et al.*, 1997) and Russia (Belaya; PROUST *et al.*, 1998) was a foredeep basin at the front of the Ural chain; it merged southward with the Precaspian basin. Facies are storm deposits (Aidaralash) and turbidites (Belaya). Its thickness is 1500 m in Aidaralash from Asselian to Artinskian. Palaeoenvironments were platform, slope and basin. Compression to extension, tectonic loading and flexural roll-back of the East European platform margin are shown by the rapid foredeep subsidence, and the westward migration of the foredeep turbidite depocentre occurred with time (CHUVACHOV & CRASQUIN-SOLEAU, 1999).

IV.4.- Precaspian basin (deep marine, turbiditic)

In Kazakhstan (ENSEPBAEV *et al.*, 1998; CHUVACHOV & CRASQUIN-SOLEAU, 1999), the Precaspian basin and its eastern border presented deep basin deposits from Asselian to Artinskian. The facies in Janajol are turbidites and in Akjar black shale and turbidites. The thickness is 500 m in Janajol and 700 m in Akjar. The palaeoenvironments were a slope in Janajol and a deep marine basin in Akjar. Deepening and drowning of Moscovian platform areas (such as Janajol) suggest continuing basin extension and subsidence. The increasing sedimentation rate is consistent with the closure of the Uralian ocean and the channelling of detrital supply from the Uralian orogen to the Precaspian basin through the Uralian foredeep.

IV.5.- Russian platform sea (shallow marine and lagoon)

Asselian to Artinskian deposits were observed in Gubakha (IZART *et al.*, 1999) near the border of the central Ural basin. The facies are bedded limestones and build-ups by algae and *Palaeoaplysina* from the Asselian to the lower part of the Artinskian, turbidites during the upper part of the Artinskian. The thickness of the limestones is 640 m and that of the turbidites more than 500 m. The palaeoenvironments were inner, mid and outer ramp with small build-ups and basin. Near the border of the Southern Ural basin (Tratau - Shaktau), thick build-ups by Bryozoa, *Palaeoaplysina, Tubiphytes* and algae formed a belt on the platform (CHUVACHOV, 1993; VENNIN, 1997; CHUVACHOV & CRASQUIN-SOLEAU, 1999).

Asselian, Sakmarian and possibly Artinskian deposits existed also in the western part of the Russian platform west of Moscow. However, the Artinskian erosional edge is found nowadays 750 km east of Moscow. The facies are bedded limestones, evaporites and claystones. The thickness is low. The palaeoenvironments were lagoon, inner and middle ramp. Tectonics: weak subsidence mainly compensated by sedimentation continued.

IV.6.- Donetz rift basin (shallow marine to paralic)

Sections were chosen near Donetz town (IZART *et al.*, 1998) in an intermediate area between continental and marine deposits from Asselian to Artinskian. The facies near Donetz are limestones, evaporites (salt, anhydrite) and red claystones during Asselian and Sakmarian and red conglomerate and sandstone during Artinskian (because of its importance, the widespread evaporite facies representing mainly the Asselian to Sakmarian condition is reported in the map, taking also into account the Artinskian erosion and fresh-water leaching of evaporites). The thickness is 1100 m near Donetz town. The palaeoenvironments were river, lagoon, inner and mid-ramp, near Donetz town. North-westward, the Dniepr basin was more continental and, south-eastward, the Russian part of Donetz basin was more marine with limestones facies. The rifting-related extension was interrupted by inversion leading to uplift of the SE part of the Donetz basin as an effect of the ending Scythian orogen compression (STEPHENSON *et al.*, 1999, STEPHENSON *et al.*, in press; STOVBA *et al.*, 1996)

IV.7.- Caucasian and Moesian basins

A shallow marine Late Carboniferous to Early Permian Caucasian seaway is assumed (EINOR, 1965), connecting the Turan sea thin-bedded turbidites and shallow-water limestones (Aral Sea wells; CHUVACHOV & CRASQUIN-SOLEAU, 1999) with the ammonoid and fusulinid limestones of Crimea (TUMANSKAYA, 1941) and the various shallow marine facies of Moesia (YANEV *et al.*, pers. com., 1998; SEGHEDI & PANA, pers. com., 1998; PANA, 1997). A large part of this basin has been tectonically buried following the Alpine shortening in Caucasus. The Zonguldak red and green alluvial-plain deposits were located at the southern edge of this seaway as suggested

by the N-pointing sediment transport direction. Strong Euramerian floral affinity is consistent with a Laurasian location of the Zonguldak area (DEMIRTASLI, 1990; KEREY *et al.*, 1985).

IV.8.- Oslo rift basin

This basin was a graben from Asselian to Artinskian (OLAUSSEN *et al.*, 1994). The facies are conglomerates, sandstones, claystones and volcanic rocks. The thickness is more than 2000 m. The palaeoenvironments were alluvial fans and rivers. Tectonics: extension, syn-rift phase.

IV.9.- Intramontane post-tectonic Hercynian basins (continental, collapse to pull-apart)

Limnic coal basins existed in Autun (BROUTIN *et al.*, 1986), Lorraine (DONSIMONI, 1981), Saar (KORSCH & SCHÄFER, 1995), Saale (SCHNEIDER, 1996), the Czech Republic (OPLUSTIL & PESEK, 1998) and Slovakia (VOZAROVA, 1998) during Autunian and Saxonian in the Saxo-Thuringian and Moldanubian zones. The facies are bituminous claystones in Autun, Lorraine, Saar and Czech basins, and sandstone and claystone in Saale and Slovak basins during Autunian, alternations of red conglomerate, sandstone, claystone and volcanic rocks in Lodève, Lorraine, Saar, Saale, and the Czech and Slovak basins. The thickness is 1525 m in Autun, 500 m in Saale, up to 1500 m in the Czech Republic and Slovakia. The palaeoenvironments were alluvial fan, river and lake. Extensional collapse extended into the outer Hercynian zones and transtensional fragmentation involved the entire domain with discrete pull-apart development. Nearly ubiquitous and abundant bimodal volcanics and volcanoclastic rocks are related to the Early Permian rift. The decreasing thickness, in comparaison with Moscovian, suggests a corresponding smoothing of the Hercynian chain relief.

IV.10.- Iran - Anatolian basin continuing in the Hellenic - Dinaric - Carnic basin (shallow marine)

This domain was partly shallow and partly deep-marine, with the Turkish block possibly separated from the Iranian block. The Elbourz basin in Iran (JENNY-DESHUSSES, 1983) (outside the map), part of the Anatolian - Hellenic - Dinaric basins, the Carnic Alps basin (VAI & VENTURI, 1997; KRAINER & DAVYDOV, 1998; DEMIRTASLI, 1990; PAPANIKOLAU & SIDERIS, 1990) presented platform deposits from Asselian to Artinskian. The facies are alternations of limestone and claystone in Elbourz and conglomerate, sandstone, claystone and bedded limestone or build-ups in the Carnic Alps. The thickness is 80 m for Elbourz and 832 m for the Carnic Alps. Palaeoenvironments were marine carbonate to silico-clastic platforms. The remaining parts of the Anatolian basins (north and south of the Kirsehir and Menderes massifs) and of the Hellenic - Dinaric basins (south of the Pelagonian and Serbo-Macedonian massifs) presented deep-water deposits. The facies are pelitic turbidites, radiolarites, abyssal clay and cherts, and are characterised by resedimented shallow-water carbonate blocks, and olistostromes (wild flysch) hundreds to thousand metres thick (DEMIRTASLI, 1990; PAPANIKOLAU & SIDERIS, 1990; RAMOVS, 1990; BAUD *et al.*, 1991). The palaeoenvironments were bathyal to abyssal basins and rapidly drowning carbonate platform edges. This sharp facies change suggests a major rifting phase consistent and coeval with the transtensional fragmentation of the Hercynian Europe, able to replace large previous platform areas with deep basins.

IV.11.- Apennine basin

The Apennine basin is an extension of the previous domain, from which it was separated by the Adriatic peninsula, representing at that time an European promontory more prominent than the Serbo-Macedonian and Pelagonian one. Tuscany and Southern Apennine basins (PASINI & VAI, 1997) presented silicoclastic to carbonate platform deposits during Artinskian. The facies are limestones and claystones. The thickness is 10 to some tens of metres. The palaeoenvironment was a large marine platform. Sicily represented the edge of the well-known drowning Sosio carbonate platform towards the deep-marine Sicani basin (CATALANO *et al.*, 1991; DI STEFANO & GULLO, 1997). Facies are shelf to reef carbonates in the platform, nodular limestones, radiolarites, cherts and abyssal clay and turbidites (containing often large exotic platform blocks and boulders) in the basin. Thickness varies from some hundreds to a few thousand metres. Rapid subsidence and platform edge collapse resulting from submarine rifting and drifting processes in nearby areas are documented here. As for the outcrops of the Greek islands, evidences of compressional shear and deformation are frequent in the Permian mudstones here, but are absent in the following similar Triassic basinal rocks.

IV.12.- Partly inferred Oman - Levantine - Sicily oceanic sea-way (Permian Tethys)

An Early Permian oceanic rift propagation along the eastern edge of Arabia is based on the Middle Permian (Wordian) age of radiolarites, turbidites and fusulinid limestone olistostromes overlaying the Oman ophiolites (BLENDIGER, 1988; BLENDIGER *et al.*, 1990). This is consistent with the pelagic Sosio-like goniatite limestone of Kurdistan, NE Iraq (VASICEK & KULLMANN, 1988) and the pelagic to turbiditic sediments of late Early to early Middle Permian age of Crete (KRAHL *et al.*, 1986), Chios, Samos, and the clastic Trogkofel basins in the Hellenic - Dinaric belt (PAPANIKOLAU & SIDERIS, 1990; RAMOVS, 1990; BAUD *et al.*, 1991). Whether this oceanic rift propagated also eastward to the Eastern Mediterranean and Ionian Seas is a matter of discussion. For this reason the corresponding part of the map (EMC and IOC) was left blank.

The Ionian and Eastern Mediterranean Seas (except for the Eratosthenes microblock) are floored by oceanic

crust overlain by an up to 9 km thick sedimentary blanket (FINETTI, 1982; FINETTI & DEL BEN, 1986; FINETTI *et al.*, 1997; MAKRIS *et al.*, 1986; SAIONI, 1996; CANTARELLA *et al.*, 1997; GARFUNKEL, 1998). This setting is also interestingly supported by the crustal features of the Levant continental margin, where the nearly normal continental crust of Negev (35 km) is laterally replaced northward by the partly thinned crust of Judea - Samaria (25 km) and further northward by the semi-oceanic crust of Galilee - Lebanon (BEN AVRAHAM & GINZBURG, 1990). This N-S gradient suggests an E-W spreading axis, quite different from the commonly restored N-S-trending rifting and spreading directions (GARFUNKEL, 1998). Much different are the opinions concerning the age of this oceanic crust ranging from Middle to Late Triassic to Middle Jurassic, Early or Late Cretaceous, Tertiary or even Messinian, according to different authors. Some authors have suggested a Permian age, although in different geotectonic settings (BEN AVRAHAM & GINZBURG, 1990; CATALANO *et al.*, 1991; VAI, 1994; SAIONI, 1996). Young Tertiary emplacements are contradicted by the cold nature of this crust (low heat flow) and the thick sedimentary cover. Evidence of deepening in the continental margin sequences are available only during Late Cretaceous, Middle Triassic and especially Early to Middle Permian. A Permian low-spreading-rate emplacement of this oceanic crust is preferred here also because of the lacking magnetic-anomaly-related banding, which could be expected for the long-lasting Permo-Carboniferous reversed polarity (Kiaman) superchron. This dating is consistent with the fact that the Early to Middle Permian is the interval when the most common and widespread occurrence of platform deepening, drowning and collapsing is documented in the areas surrounding the Ionian and Eastern Mediterranean Seas. Also the high thickness of sediments overlying the oceanic crust, in relatively starved to partly supplied pelagic basins, requires an old age of this crust. Different attempts to trace basinward calibrated seismic profiles from the continental margins (such as those by GARFUNKEL, 1998, Fig. 2) show that pre-Jurassic sequences are considerably thickening in the basins. The above reported are some of the arguments that make the assumption of a Permian age for the Ionian and Eastern Mediterranean oceanic crust more attractive than other younger dating, unless direct documentation is provided.

Evidence of oceanic crust ceases at Eastern Sicily. However, 1700 m thick Permian basinal and platform marine deposits overlying a sialic crust are known through Sicily up to the Djeffara basin, South Tunisia. The abrupt disappearance of this marine belt is not palaeogeographically but tectonically controlled. In fact, the Djeffara beds plunge and continue in the subsurface beneath the Tunisian front of the Atlas Maghrebian thrust belt (AtMD) (VAI, 1997). The AtMD is blank on the map because the thrust sheets (except for the Rif and Kabylia teleallochthonous nappes showing continental conditions) are detached at the Late Triassic evaporite level, thus preventing any information about their original Permo-Carboniferous substratum. This means that an EW-trending Early to Middle Permian marine seaway could have existed over the blank Maghrebian area up to the Moroccan Atlantic coast. In fact, evidence of pre-Triassic (?)Permian synrift deposits is found in seismic profiles of the Agadir - Tarfaya offshore (cf. NAHIM & JABOUR, 1997).

An Early to Middle Permian shallow marine seaway connecting the Permian western Tethys to the eastern Panthalassa would fit in well with the Permian migration of Tethyan fusulinid and brachiopod faunas through Cuba into Central America (outside the map). It would also prevent the exchange of floras and continental tetrapods between Laurussia and Gondwana. In fact, the early evolutionary radiation of tetrapods suggests that continental migration routes from Laurussia to Gondwana and *vice versa* were effective only by the end of Permian (SCHNEIDER, 1996). Reptilian faunas are known from the Late Carboniferous of Laurussia (including the Carnic Alps). Records of Early Permian tetrapods from Gondwana are restricted to few mesosauroids of South Africa and South America and amphibians of Kashmir and North-Westernmost Africa (MILLSTEED, 1995). Terrestrial reptiles do not appear in the fossil record of Gondwana until the Late Permian, despite the dominance of terrestrial sediments there during Late Palaeozoic. The assumed Early to Middle Permian North African to Caribbean sea-way acted as the major barrier preventing the migration of terrestrial forms between Laurussia and Gondwana. The same marine barrier-controlled migration pattern could also explain the spreading of *Glossopteris* floral elements through NW Africa and Iberia towards Asia starting only at the Late Permian. In this assumption, marine conditions along the seaway were replaced by continental ones during the Late Permian, which is consistent with the following map of the Atlas (Wordian).

Such a model is also consistent with a transtensional dextral slip between Gondwana and Laurussia lasting some tens of millions of years, occurring coevally with the compression and shortening accomplished in the Ouachita - Allegheny - Mauretanian belt and in the Uralian one. The opening and temporary spreading of the Permo-Triassic Tethys during Early to Middle Permian explain this setting. A first activation of the Somaliland-Tanzania rift probably stopped the emplacement of oceanic crust in the Permo-Triassic Tethys (including the Eastern Mediterranean and Ionian areas), leading to a transpression with uplift in the Maghrebian corridor, and completed the compression in the Allegheny - Mauretanian and the Uralian belts during Late Permian.

IV.13.- South Peri-Tethyan platform basins (shallow marine and continental)

The Moroccan basins (EL WARTITI *et al.*, 1990) were continental during Autunian and Saxonian. They were still located in a North American position, north of the assumed Maghrebian-offshore Moroccan seaway. The Algerian basin of Kenadza (DELEAU, 1951) was continental during Autunian. The Tunisian basin (Kirchaou; LYS, 1988) was marine from Asselian to Artinskian (see above), same as the Cyrenaic basin (VACHARD *et al.*, 1993) during Early Permian. The NE Egypt basin (KORA, 1998) was marine from Asselian to Sakmarian and continental during Artinskian. The Bademli basin in South Turkey (MONOD, 1977) and Syria basins (AL YOUSSEF &

AYED, 1992) were marine during Early Permian. The facies are alternations of red conglomerate, sandstone, claystone in the Moroccan and Algerian Basins. In Tunisia, the facies were limestones and marine claystones and sandstones, claystones or limestones in Libya, limestones in South Turkey and marine claystones and sandstones in Egypt and Syria. The thickness was 3000 m in Morocco, 50 m in Kenadza, 600 m in Tunisia, 437 m in Cyrenaic (Libya), 192 m in Egypt, 40 m in Southern Turkey and 600 m in Syria. The palaeoenvironments were alluvial fans and rivers in Morocco and Algeria, continent or marine platform in Tunisia, Libya, Egypt, South Turkey and Syria. Middle Permian (Wordian) 4000 m thick turbidites are known from the Tebaga well in Tunisia (LYS, 1988) (see above). Extension to transtension-related subsidence dominated this domain.

IV.14.- Cratonic East Arabian basin (continental, lagoon and platform)

Saudi Arabia (OWENS & TURNER, 1995; AL LABOUN, 1993) and Oman (LOVE, 1994; BROUTIN *et al.*, 1995; ANGIOLINI *et al.*, 1997) presented continental and marine deposits during Asselian to Artinskian. The facies are sandstone and claystone during Asselian and Early Sakmarian and sandstone and limestone during Late Sakmarian and Artinskian. The thickness is 400 m in Saudi Arabia and 100 m in Oman. Palaeoenvironments were rivers and alluvial plains, and a marine platform. Early Permian passive continental margin extension and subsidence is shown by the transgression reflecting the Permo-Triassic Tethys spreading.

IV.15.- Intracratonic North African basins

Continental conditions extended to all basins characterised by thin and discontinuous Early Permian deposits. Facies are mainly sandstones and sandy claystones, associated with volcanoclastic rocks in the Taoudenni basin (LEGRAND-BLAIN, 1985). The thickness is 70 m for the Ghadames (Libya) basin (COQUEL *et al.*, 1988; MASSA & VACHARD, 1979) and possibly 150 m for the Illizi basin (LEGRAND-BLAIN, 1985). E-W-trending normal faulting seems to control deposition of the Erdi basin, in a picture showing intraplate deformation and anorogenic alkaline magmatism in NE Africa and Near East, producing transtensional basins and regional uplifts with dextral sense of displacement (SCHANDELMEIER *et al.*, 1997).

3.- WORDIAN (266 - 264 Ma)

M. GAETANI[1]

with contribution of
B. CHUVACHOV

I.- MAIN FEATURES

The Permian stratigraphic subdivisions have been undergoing a significant change and upgrading in the last few years. Traditionally, the Permian was subdivided into two series. In continental Europe, the Rotliegend was considered as representative of the Early Permian, whilst the Zechstein was thought to represent the Late Permian. In the Tethyan or Uralian marine dominated environments, the basal boundary of the Late Permian was drawn at the base of the Kubergandian or Ufimian, respectively. In 1997, the International Subcommission for Permian Stratigraphy divided the Permian in three series (JIN *et al.*, 1997). Therefore, the selected time slice of the present map, formerly considered as belonging to the Late Permian, now falls within the upper part of the Middle Permian.

The present map corresponds to the first map of the Tethys Project (BAUD *et al.*, 1993), in which the time slice adopted was the Late Murgabian, thought to correspond to the fusulinid *Neoschwagerina margaritae - Eopolydiexodina* zone, and to the conodont *Neogondolella aserrata* and *N. postserrata* assemblage zones. The Murgabian, a Tethyan stage, was defined in South Pamir (MIKLUKHO MACKLAY, 1958; LEVEN, 1967). The assemblage *N. margaritae /Yabeina* was later recognised to be correlative of the subsequent Midian stage of the Tethys. The stage Wordian is partly overlapping the Murgabian and is here preferred because of the common proposal of the ISPS (JIN *et al.*, 1997) and the CGMW (1998). Also, even if the use of Guadalupian series is still debated, the Subcommission for Permian Stratigraphy of the IUGS, has recently approved its use, with three stages, Roadian, Wordian and Capitanian in ascending order, with type sections in Texas, USA (GLENISTER *et al.*, 1999; WARDLAW *et al.*, 1999). We consider here the Late Wordian as represented biostratigraphically in the Tethys by the conodont *aserrata* zone and by the fusulinid *Neoschwagerina margaritae/Yabeina archaica* (now early Midian in the Tethyan scale (DAVYDOV, 1996). However, and beyond nomenclatural problems, the correlation with marginal or continental areas of the Peri-Tethyan realm is difficult on pure biostratigraphic basis. The magnetostratigraphic approach introduced a new powerful tool, especially as regards the Kiaman/Illawarra reversal and the base of the Illawarra mixed polarity zone that seem to approch in the Salt Range, Pakistan, the base of the Late Murgabian (HAAG & HELLER, 1991). In the Nammal Gorge (HAAG & HELLER, 1991), the K/I is not recorded. The sequence of polarity reversals starts with a base-lacking normal polarity zone of Late Murgabian age. These authors conclude that the K/I boundary must be older than Late Murgabian. The correlation between the Nammal Gorge magnetostratigraphy and the Pre-Ural basin magnetostratigraphy (GIALANELLA *et al.*, 1997), the latter containing most probably the K/I boundary, is visibly convincing and suggests that the K/I boundary in the Nammal Gorge should be located immediately below the base of the section. This reversal has been used to draw the correlations in the present map. In the Pre-Ural basin the K/I reversal falls within the Tatarian beds (at the boundary between Urzumskian and Severodvinskian Svita) (GIALANELLA *et al.*, 1997), ruling out the correlation Murgabian = Kazanian. In the Middle European basin the reversal lies in the lowest part of the Upper Rotliegend (MENNING, 1995) ruling out the usual correlation Murgabian = Zechstein. This is the reason why the present palaeogeographic map appears to be so different in comparison to the previous Tethys map. JIN *et al.* (1997) set the K/I boundary within the Wordian. Afterwards, however, doubts arose about the actual position of the reversal, since it might also lie around the boundary Wordian/Capitanian or even within the basal Capitanian (GLENISTER *et al.*, 1999; MENNING, pers. comm., Febr. 2000). The K/I boundary is used as the main correlation tool in the present map, whenever it is known. In continental successions like in France and Spain, where the magnetostratigraphic control is not available, the correlation has been tentatively made through the "Saxonian floras". In the marine successions of the Arabian Peninsula, where the magnetostratigraphic control was not available, the biostratigraphic tool was used. The best radiometric age for the top of the Wordian

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is presently around 265 Ma (WARDLAW, 1999). The base of the overlying Capitanian has delivered an age of 265.3 ± 0.2 Ma (U-Pb isotope dilution mass spectrometry on zircons) (BOWRING *et al.*, 1998). The time scale proposed by ODIN (1994) doesn't appear to be reliable for the Middle and Late Permian. To conclude this discussion, the selected time slice corresponds, where possible, to the Late Wordian, equivalent to the Late Murgabian of the Tethys project (BAUD *et al.*, 1993), now Early Midian in the updated Tethyan scale (DAVYDOV, 1996).

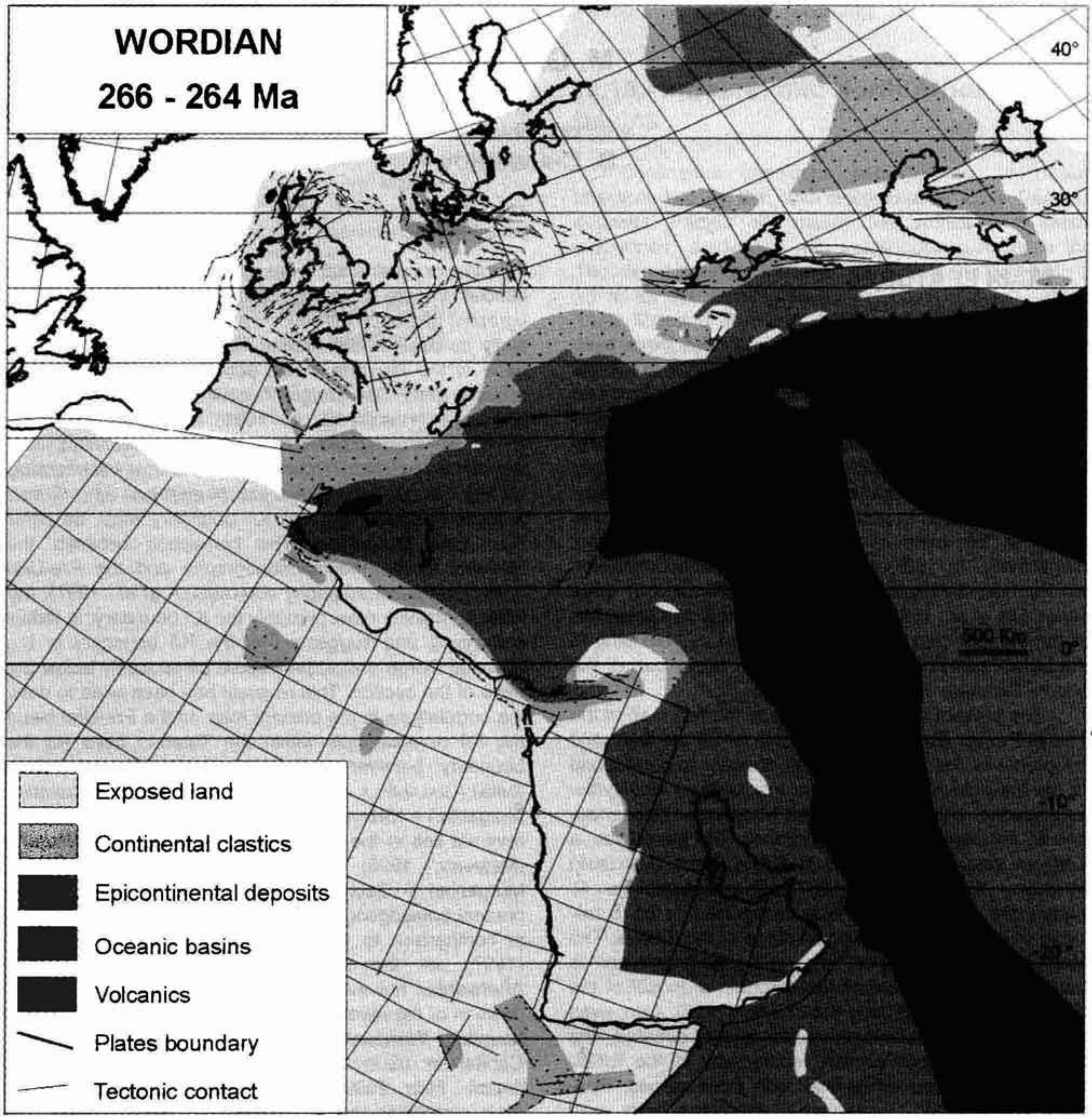

Fig. 3.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Wordian.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

As already discussed in the general features of the Pangaea times, three main kinematics trends have developed during the Late Permian:

1.- the opening of Neo-Tethys, with an anti clock-wise rotation of the Cimmerian blocks and consumption of the Palaeo-Tethys along a northward subduction zone, active to the south of the present Caucasus and Kopet Dagh Mountains;

2.- a dextral transcurrent regime between Laurussia and Gondwana; and

3.- the final stages of the convergence between Laurussia, Kazakhstania and Angara, resulting in the Ural Orogenic belt.

The main continental blocks accounted for in this time are:

- the East European platform (EEP) in which the Pripyat, Dniepr, Donetz, Karpinsky alignment of basins experience a significant lowering, even a stop, of its subsidence, before the Late Triassic inversion;
- the system of troughs and basins that border the East European platform and the Baltic Shield from the North Atlantic rift to the Caucasus, via the Polish Trough and North Dobrogea;
- the collage of minor blocks that will result in the Turan Plate with the Cimmerian orogeny;
- the microblocks of European affinity, like Moesia, Istanbul, East Pontides, Dzhirula that line the hanging wall of the subduction zone of Palaeo-Tethys;
- the Western Europe block, in which an extensional regime with a significant N-S rifting is coupled with the pull-apart basins opening along the transtentional regime active at the border of the Baltic Shield and East European platform;
- the Iberia Block, separated by two transtentional alignments from Western Europe and the northern rim of Gondwana in Africa;
- the Gondwana margin, mostly with passive continental margin behaviour, except for the Maghreb segment.

II.2.- Palaeoposition of plates and blocks

No new palaeolatitude data have been introduced in the present map (Fig.3.1) . VAN DER VOO (1993), MUTTONI et al. (1996), TORCQ et al. (1997) and BESSE et al. (1998) have discussed the existing data and Apparent Polar Wandering (APW). Palaeolatitudes obtained by LEMAIRE et al. (1998a and b) on Permian rocks in Turkmenistan, by FEINBERG et al. (1996) on the basal Triassic of Mangyshlak, Kazakhstan, as well as those calculated by G. MUTTONI (pers. comm., 1998) based on data of GIALANELLA et al. (1997) concerning the Tatarian succession of Pre-Urals, have not been adopted in the present map. Beyond the poor stratigraphic control of LEMAIRE's data, all the calculated palaeolatitudes are too low and do not fit in neither with the climatic pattern that can be inferred for the Transcaspian area (GARZANTI & GAETANI, 2000) nor with the general palaeogeographic picture obtained from the APW of major plates. The diagenetic squeezing of the red pelites and siltites from which most of the samples have been obtained, could perhaps explain such a bias.

II.3.- Accuracy

The inferred dextral transcurrent movement on the southern rim of the East European platform took place widely along lines of latitude. Given the geocentric axial dipolar nature of the palaeomagnetic field, no longitudinal constraint on the motion can be directly obtained by palaeomagnetic analyses. The accuracy of a good palaeolatitude datum is typically in the order of a few degrees, corresponding to a few hundreds kilometres. Therefore, small geologic features like those which are supposed to have formed during the transtensional regime, which characterised the Wordian cannot be confidently located with the exclusive use of the palaeomagnetic method. In the Turan area, the Karabogaz High was moved to the south with respect to the previous version of the map to account for back arc extension which occurred to the north of it. This back-arc basin was later tectonically compressed during the Eo-Cimmerian orogeny. However, the amount of shortening required by the LEMAIRE et al. (1998a and b) model is here considered too large and not supported by geological data. The Moesia Block was moved towards the south-east to allow the shallow water fusulinid sea arm that has been detected in Central Moesia (PANA, 1997) to be connected to the Palaeo-Tethys ocean.

III.- DEFINITION AND DESCRIPTION OF THE DOMAINS

III.1.- East European platform

This wide area forms one of the largest domain considered during the project. Three subdomains will be taken into consideration: the Pre-Uralian (Preduralian) basins, the Precaspian basin and the Pripyat - Donetz basins. The southern and south-western margins will be studied separately. The Baltic Shield extended eastwards with two major positive structures, namely the Voronezh High and the Ukrainian High and in between, the NW-SE aligned sequence of the intracratonic basins Pripyat - Dniepr - Donetz. The inversion of the Karpinsky - Donetz foldbelt sections started during the Early Permian and the transition to uninverted occurred at the western tip of the Donbass foldbelt (STOBVA & STEPHENSON, 1999). Details of the basins evolution are analysed in STOBVA et al. (1996) and VAN WEES et al. (1996).

The Late Permian sediments of the Donbass are the residual infilling of the post-rift succession that reached a maximum thickness of few tens of metres. Even their actual age assignment might be argued (STOBVA et al., 1996)

III.1.1.- Pre-Urals (B. CHUVACHOV)

This system of basins extended from the present Arctic sea to the steppes of the NW Kazakhstan over 2500 km in length. It represents the flexural margin of the EEP to form the fore-arc basin of the Ural Orogen. Largely marine during the Early Permian, it gradually developed a marginal to continental facies with short living marine ingressions. The last generalised ingression occurred during the Kazanian, the stage preceding the Tatarian in the Uralian scale, which is here considered partly equivalent to the Wordian. During the Tatarian; there was still a strong subsidence (about 1800 m of continental terrigenous sediments in the southern part of the basin, and up to 2300 in the northern part (CHUVACHOV et al., 1984, 1995) and the marine connection was still opened to the north in the Pechora basin area (East European Sedimentary basin) (MURAVYEV, 1972; TIKHOMIROV, 1984).

A significant brackish-water basin extended from the Kanin Peninsula up to present latitude 55 N. It could be considered as the Kanin - Vychegda Bay of a very large Tatarian Sea, which occupied the present territory of the Barents Sea. More deep-water part of Kanin - Vychegda Bay is traced along the Timan land. Here sediments consist of marls and argillites with subordinated sandstone and limestone in thin layers and packets. The

biota of this part of the Tatarian basin was rich and diverse, including foraminifers, bryozoans, and calcareous algae. The general thickness of the Tatarian stage within the Kanin - Vychegda Bay is of 200 - 500 m.

A very low relief hilly land was bordering the eastern part of the basin. Fluvial plains to the west and to the south show evidence of the belt of mostly red-coloured terrestrial sediments (clays, silts, sands and rare evaporites). Coarser sediments were deposited near the Ural Mountains areas. The first of these belts is situated between the Northern Urals and the Timan lands, where grey and red-coloured terrestrial sediments accumulated. Locally they contain coal seams. The thickness of the whole Tatarian could locally reach 2300 m. A second area of coarse-grained sediments might be traced along a semioval belt, from Orenburg to Astrakhan. Tatarian sediments are represented here by red-coloured sandstones, silts, argillites. To the east of this belt conglomerate lenses may also occur. Locally horizons of the grey-coloured sediments with numerous stromatolite bodies are present. Moving to the west, the coarse sediments grades into red-coloured siltstone and claystone. The large area occupied by the sediments of Tatarian age, located between Orenburg and Astrakhan, can be named as the Volgo-Urals fluvial plain.

Tectonic block movements occurred during the Tatarian on the territory of the Urals. The results of this process was a narrow and long system with N-S oriented depressions (CHUVACHOV *et al.*, 1984).To the south, the thickness of the Early Tatarian, is around 100 m on the sill between the EEP and the Pre-Uralian fore-basin, whilst in the basin itself it may reach 800 m in thickness (MOLOVSTOSKY *et al.*, 1998). The dominant facies are red shales and siltites, with subordinate conglomerates.

III.1.2.- Precaspian

The whole Precaspian basin continued to produce large accommodation room for sediments after the salt deposits of the Kungurian (= base of the Guadalupian Series). Outcrops are present only around the northeastern rim of the basin, from Orenburg to Aktyubinsk, in the transition from the southernmost part of the Predural and the Precaspian depression. Here, the two "Svita" of the Early Tatarian are about 1000 m thick, presenting a especial dominion of shales in the middle upper part. Facies indicate large alluvial plains and playas, mostly fed from the Ural Orogen, as suggested by the coarser deposits towards the north-east (MOLOVSTOVSKY *et al.*, 1998). In the centre of the basin thickness is still relevant and the whole package between the salt deposits and the Early Triassic marine ingression is between 2 and 4 km. Detailed thickness for the Wordian cannot be always indicated, depending on the availability of deep boreholes. Thickness is also controlled by later local erosion due to uplifts linked to the salt diapirism. The total thickness of the Tatarian may reach 2030 m, though the Early Tatarian is usually around 1000 m thick (KUKHTINOV & CRASQUIN-SOLEAU, 1999).

III.2.- Border basins at the margin of the EEP

III.2.1.- "Scythian platform" and Caucasus

This part is often considered as an independent crustal block from the EEP (NIKISHIN *et al.*, 1996, 1998a). Largely concealed under the alluvial plains, the southern boundary of the EEP is drawn at the northern margin of the Hercynian deformation, where the strike-slip system of faults allowed the movements of the mobile Caucasus. No data are known for the Kuban and Terek basins, possibly not yet identified at that time. Permian deposits are preserved in the western part of the Caucasus Range, whilst in the eastern part, if existent, they were eroded away during the Cimmerian orogeny. Very thick continental deposits are preserved inside tectonic grabens and part of them could be of Late Permian age. No palaeomagnetic stratigraphy is known to allow the Kiaman/Illawarra boundary recognition. Two major areas with red conglomerates and feldspathic litharenites to lithic arkoses of Verrucano-type are indicated tentatively in the Great Laba and Belaya river areas, where they may reach 2000 m in thickness, as a whole. In general, rhyodacitic volcanic pebbles do not exceed the 10% of clastics (GARZANTI, pers. comm., 1999). The marine ingression occurred slightly later, during the Dorashamian, according to the Tethyan scale (KOTLYAR, 1998). Several adjacent blocks are devoid of Permian deposits and probably fed the grabens with sediments.

III.2.2.- Crimea

No data are available for the Permian in areas of the northern and central Crimea.

In the south, in the Gornyi Crimea, limestone blocks containing Wordian fauna of Tethyan affinity have been known since a while (FOKHT, 1901; TOUMANSKY, 1941) embedded in the Tauric Flysch. Within the PTP, two projects have been funded to review the stratigraphy of these blocks (KOTLYAR *et al.*, 1999; GRUNT, 2000). These shallow water limestones are interpreted as deposed in the shallow back arc basin lying to the north of the Palaeo-Tethys subduction margin.

III.2.3.- North Dobrogea

In North Dobrogea numerous grabens are recognised, mostly located to the north of the Sfintu Georgi Fault in the Lower Danube Graben. The clastic infilling may reach several hundreds meters for the whole Middle - Late Permian. However, no certain age assignments are available. Erosion continued on the horsts. Volcanic activities with alkaline flood rhyolites within intraplate signature are also reported (SANDULESCU, 1984; SEGHEDI & OAIE, 1995; SEGHEDI, 2000). The large amount of synrift-magmatic rocks, including alkali granites and syenites associated with alkali rhyolites was emplaced along the southern margin of the North Dobrogea (SEGHEDI, 2000). Syn-rift evidence is also provided by the alkaline dyke swarms along the Sfintu Georgi Fault, which are unconformably overlain by the Early Triassic conglomerates (SEGHEDI, 2000).

III.2.4.- Polish Trough

The Saalian transpressional event led to a general inversion of the sedimentary basin of Lower Rotliegend as the Wordian time was largely a period of erosion. Only limited areas along the Polish Trough and in the northeastern part of the Middle European basin constituted sites of deposition for the Upper Rotliegend that eventually evolved later in the Zechstein basins. The map was drawn at the level of Parchim and Drawska formations. They consist of conglomerates, sandstones and siltites in the centre of the depressions, with playa sediments assemblages. The WNW-ESE aligned basin

formed large intramountane depressions. Locally, also eolian sands are reported testifying to the very arid environment (McCANN, 1998; KIERSNOWSKI, 1991; POKORSKI, 1997). The main basin was lined by minor depressions with the same alignment, which concurred to form the initial part of the Middle European basin in the area (SCHECK & BAYER, 1999). These basins and depressions were most probably activated and lined by extensional or transtentional faults. It is difficult to assess whether they were actually active during the selected time slice of the map or represented a general trend of the whole late Middle to Late Permian.

III.3.- Turan area

The region east of the Caspian Sea is usually recorded in the literature as Turan Plate (VOLVOSKYI *et al.*, 1966; MUROMCHEV, 1968). However, the final consolidation of the Turan Plate occurred only as a consequence of the Cimmerian Orogeny in the Late Triassic. Four main areas are identified: North Ustyurt, Mangyshlak, Tuarkyr and Karakum. The North Ustyurt and the Karakum were apparently emerged areas subject to erosion (COOK *et al.*, 1994; GAVRILIANSK, 1965). In the Gornyi Mangyshlak, the base of the outcropping succession was often considered of Permian age (MUROMCHEV, 1968: LICHAREV *in* STRATIGRAPHY OF THE USSR, 1973), but palynological dating has demonstrated that only Early Triassic rocks are present (LOZOWSKI *et al.*, 1986). In the subsurface of the Southern Mangyshlak, possible Permian continental deposits are present, but the recent revision of the area by THOMAS *et al.* (1999), mostly based on a reassessment of Soviet data, consider the succession, as a whole, as the "Intermediate sequence" of Permo-Triassic age, without further subdivisions.

In the Tuarkyr, continental sediments deposed under a semiarid and arid environment with flash-flood streams containing abundant volcanic pebbles, tuffs of felsic composition, as well sedimentary pebbles, eroded by a nearby emergent carbonate platform with numerous foraminifera, sponges and hydrozoans of Devonian age (GARZANTI & GAETANI, 2000). A tentative Late Permian age is attributed to this succession.

Volcanic activity with effusion of dacitic-rhyodacitic lavas should have been significant at the southern margin of the Karabogaz High (Krasnovodsk - Turkmenbasi area) (LEMAIRE, 1997). However, precise timing is not available.

III.4.- Marginal sea and microblocks of European affinity

Along the southern margin of the European Plate, there are several crustal fragments of European affinity that are now largely included in the Alpine orogenic systems, but Moesia. They are interpreted in the map to lie between the subduction front of the Palaeo-Tethys Ocean and the transtentional basin belt that bound the East European platform and the Turan area. They may also form the sialic core of the volcanic arc growing on the hanging wall of the subduction zone. The interposed sea is interpreted as the back arc basin linked to the roll-over effect of the arc.

Remnants of the back arc sea are scanty and could be, from east to west:

- the Dizi terrigenous succession. On the southern slope of western Caucasus a flyschoid succession is considered to span from the Permian to the Early Jurassic (SOMIN & BELOV, 1967);
- the exotic blocks of the southern Crimea, previously described;
- the sedimentary fragments floating over the Küre Complex (USTAOMER & ROBERTSON, 1997).

At the southern margin of the region, forming the core of the arc, there are several crustal blocks. They are, from east to west:

- Dzhirula and Khrami in the Transcaucasus, slabs strongly deformed during the Hercynian Orogeny with evidence of intrusive and sedimentary rocks of Palaeozoic age, probably emergent during the Late Permian (ADAMIA & LODKIPANIDZE, 1989);
- the Eastern Pontides, in which Early Carboniferous "Granodiorite - Dacite Complex" and Late Carboniferous - Early Permian? "Sedimentary Sequences" are recognised. The latter contains about 1000 m of monotonous red silicoclastic sandstones possibly of Permian age (OKAY & SAHINTÜRK, 1997);
- the Central and Western Pontides, where the Istanbul Zone or Unit is preserved (OKAY, 1989; USTAOMER & ROBERSTON, 1997). No Permian sediments are known for sure from this unit, but the lithological affinities of the Palaeozoic sequences testify to a northern origin of the zone itself (SANDULESCU, 1978; OKAY *et al.*, 1994; BANKS & ROBINSON, 1997). Tentatively, the Tepekli conglomerate could be attributed to the Permian (S. DERMAN, pers. comm., 2000);
- the Moesia. The interpretation of the palaeoposition of the Moesia Block and its geodynamic significance is one of the major differences in comparison to the Tethys map (BAUD *et al.*, 1993). The reasons are two fold. The analysis of a number of cores from drillings on the Romanian side of the Moesian platform evidenced the presence of a marine gulf with fusulinids in the Middle and Late Permian (PANA, 1997). The present position of this marine gulf in the middle of Moesia against the North Dobrogea, oblige to find a connection with the open sea. The southern connection is here excluded because both the Pre-Balkan and the Balkan itself were sites of important continental erosion. Directly to the east, the drillings have only recorded, up to now, non-depositional conditions during the Late Permian or small grabens with continental clastic infilling or evaporite deposits. Apparently, the only possible connection is towards the north-east. To have room for such a connection, it would have been necessary that the Moesia Block move consistently either to the south in order to have a hypothetical seaway along the southern rim of Dobrogea grabens, or towards the SE of at least 200 km, in order also integrate the Jurassic - Cretaceous dextral transpression along the Peceneaga Fault (SANDULESCU, 1990). Worth to be noted is the fact that this solution is opposite to the BANKS & ROBINSON's (1997) one, that locates Moesia towards the NW at the beginning of its history. However, this configuration makes much more difficult finding a connection with this fusulinid sea in the middle of the emerged Moesia Block;
- the Rhodope. Without entering in the controversies concerning Rhodope (ZAGORCHEV, 1998), we assumed that at least part of this crustal fragment was lying on the hanging wall of the subduction zone of the Palaeo-Tethys, in agreement with USTAOMER & ROBERTSON (1997).

III.5.- The Mid European basin and Western Europe

As consequence of the Saalian convergence event, most of the Western Europe was emergent and under erosional conditions. Temporary alluvial deposits were accumulating along the depressions originated by the strike-slip tectonic system, but few of them are preserved or correlatable for sure to the Wordian time. No magnetostratigraphy of these deposits is available and the so-called Saxonian successions have been tentatively considered of Wordian age in the present map. Only in the initial development of the Mid European basin the magnetostratigraphic signature allows a correct correlation (McCann, 1998). The internal depressions are aligned with the Polish Trough, with the distribution of the depocentres clearly controlled by a pull-apart basin system (McCann, 1998, figs 6, 7). In the collapsing Hercynian complex, the preserved grabens are mostly oriented N-S in the rifted segments, usually 10 to 30 km wide and at least 100 - 200 km long. They are present on the margin of the Bohemian massif (Pesek *et al.*, 1998) as well in the central part of the North Sea (i.e. Horn Graben). However, as it may be recognised also from the successive Zechstein depocentres, the general trend of the basin is WNE-ESE (Geluk, 2000).

Isolated graben structures hosted the early post-orogenic deposits in Western Europe like the few alluvial deposits known around the Massif Central in France (Lodève, St. Affrique) (Broutin *et al.*, 1992, 1994) or the Exeter Group in south Devon, British Isles (Edwards *et al.*, 1997). Documentation of the tectonic activity is difficult due to the limited preservation of these early infillings of the grabens. However, the general trend towards the collapse of the Variscan orogenic build-up during the Middle/Late Permian may postulate the introduction of several tectonic lineaments in the map, even if they are difficult to prove.

Magmatic activity is strongly reduced during this time in this part of Europe.

III.6.- Iberia

Interpretation of the clastic sequences in Iberia evolved in the last years. Sopena *et al.* (1988) recognised the "Saxonian" in Molina de Aragon rifted basin as well in the Lerida area of Pyrenees. Instead, Lopez-Gomez & Arche (1993) and Arche & Lopez-Gomez (1996), on the ground of the sequence interpretation, referred no more sediments to the Saxonian.

Therefore, a Thuringian age (European continental stage equivalent to the latest Permian) is preferred instead for the sequence No. 2 and 3, traditionally called "Saxonian". Adopting the most recent interpretation, no Wordian-equivalent sediments are recognised in Iberia.

III.7.- Gondwanan margin

III.7.1.- Maghreb

No Late Permian sediments are reported in Morocco (El Wartiti, 1996) and Algeria. Instead, in the south of Tunisia the Permian outcrop of the Tebaga of Medenine is well known. It is a part of the Permian engulfment well constrained by several boreholes and seismic lines. The southern shore of the Wordian sea was about at the present parallel N33°, roughly trending E-W. The shallow water shaly to carbonate complex is about 60-80 km wide, to plunge to deeper shaly facies towards the present Mediterranean. This is the most extended encroachment of the Tethys sea on this part of the Saharan platform during the Permian.

III.7.2.- Libya and Egypt

At the south and east of Tunisia, the Gondwana shores laid to the north of the present North African continent during the Middle and Late Permian. Consequently, the whole Libya and Egypt were in a continental and mostly erosional setting. Moreover, the preserved sediments in the internal basins like the Murzuk, Al Kufrah and Syrte basins in Libya or the Wadi Araba in Egypt, contains the so called Nubian Sandstone Group, which may span from the Late Permian to earliest Cretaceous. Consequently, the presence of Middle - Late Permian sediments in the map is only tentative.

III.7.3.- Levant

The Near and Middle East have been extensively analysed recently by Alsharhan & Nairn (1997).

Here we summarise only the major points of the map. The Middle and Late Permian rocks in the Near East are better expressed in the Palmyra Embayment, an elongated depression activated by extensional faulting, more significant to the south. Outcrops are generally missing and almost all the data originate from borehole. To the south, the Palmyra Embayment is bounded by the Rutbah High, which has no Permian and Triassic deposits. However, its emersion during the Wordian is only assumed, since the Rutbah High was active during the Jurassic and consequently the Permian and Triassic sediments could be also missing because of later erosion. The same holds true for the northern part of the Saudi Arabia margin, because of the subsequent erosion of the pre-Cretaceous sediments to the north of the 28° parallel.

As the accuracy of datings is usually poor, the solution proposed in the map in order to have Wordian shallow water carbonates in the centre of the Palmyra Embayment is in fact a compromise between Sawaf *et al.* (1993) and Alsharhan & Nairn (1997, fig. 6.22). The former found mostly shales and marls in the transect along the 40° East longitude within the Amanous Formation, as well as along the NW oriented transect between Palmyrides and Aleppo Plateau (Best *et al.*, 1993). Alsharhan & Nairn (1997), for their part, following Al Youssef & Ayed (1992), reported more than 200 m of lime- mudstones in the Palmyra Embayment for the Late Permian (the Middle Permian is not distinguished in these papers). Also, de Ruiter *et al.* (1994) don't quote a prominent carbonate level in the Permian. The not well constrained dating hampers a detailed palaeogeographic map for an area where shallow water mudstones, shore-face clastics and continental alluvial plains are interfingering. The same trend is present to the south, in the marginal area of Negev (Israel) and in Sinai (Egypt). In the borehole Maqtesh Daran 2 (Negev), calcareous inter-calations with fusulinids of Late Permian age were also penetrated (Garfunkel & Derin, 1984).

The Aleppo Plateau and the Mardin area were emergent at that time and the Syrian promontory was elongated towards the north by a wide structural platform inundated by the shallow water carbonates and clastics, later involved by the Arabia/Asia collision to form the Tauride Range. The interpretation of this area is beyond the scope of the PT Programme, through the map is

drafted according to the ideas of A. POISSON (Orsay Univ., France, pers. comm. 1999) (OZDEN *et al.*, 1998).

III.7.4.- Arabia and Oman

The Arabian Peninsula is characterised by the large epicontinental sea, which gradually aggraded on the flat margin. This belt has an average of 1000 km, from the rift shoulder facing Neo-Tethys to the shores against the Arabian Shield. Because of this aggradation, a complete set of gradual transition is present, from open ramp carbonates, via the shallow water carbonate banks and sheltered evaporitic lagoons, and subsequently from facies with variable salinity to end in the shore-face and alluvial plain terrigenous sediments. The aggradation was heterochronous and this is reflected in the age of the base of the most important marine unit, the Khuff Formation. On the marginal area in Saudi Arabia, the clastic facies are distinguished as the Unayzah Formation. Due to its importance in the Permian petroleum system, the Khuff Formation has been studied in the last year quite extensively and subdivided in several sectors, which have been correlated mostly by means of physical stratigraphy tools (KING, 1995; LE MÉTOUR *et al.*, 1995; AL JALLAL, 1995; LE NINDRE *et al.*, 1990a) and summarised by AL ASWAD (1997) and ALSHARAN & NAIRN (1997). Age assignments are often of limited reliability in the details, and were done mostly using the Uralian classification of Kazanian and Tatarian, but without magnetostratigraphic control. Unsuccessful attempts were made within the Peri-Tethys Programme to measure magnetostratigraphy in the Late Permian of Oman (J. MARCOUX, Paris, 1996, pers. comm.). Only in few places like in the Haushi area (Oman), the Wordian can be identified by means of conodonts and brachiopods (ANGIOLINI *et al.*, 1998; ANGIOLINI & BUCHER, 1999) and ostracods (CRASQUIN-SOLEAU *et al.*, 1999). Actually, thanks to frequent drillings carried out in the Khuff Formation and its equivalents, the knowledge of the companies on the Oman Mountains and the margins of the wide continental submerged and emerged platforms, where outcrops are widespread, is wider than the data published.

The margin towards the Neo-Tethys has been largely studied in the Oman Mountains. Also significant is the Zagros High (SZABO & KHERDAPIR, 1978) that could be interpreted as an emergent rift shoulder, on the border of the passive margin facing the Neo-Tethys. The basic volcanics poured out in the Hamrat Duru basin (Oman) (LE MÉTOUR *et al.*, 1995) are also of interest.

III.7.5.- Somalia and Ethiopia

Data are scanty. Permian sediments are interpreted as part of the pre-rift infilling of the tri-radial system of grabens, including Abai River and Ogaden (BOSWORTH, 1994). In the Ogaden basin, the base of the Karroo-like sediments (lower Bokh Shale above and Calub Sandstone below) were penetrated by drilling, especially in the Calub area. The lower Bokh Shale consists of dominant dark grey shales and silts, with Estherians, bivalves and fishes of lacustrine environment, whilst the Calub Sandstone contains coarse arkosic sandstone with rare intermediate to basic tuffitic intercalations (WORKU & ASTIN, 1992). Their age is roughly referred to the Middle and Late Permian (HANKEL, 1994; HUNEGNAW *et al.*, 1998). In the centre of the basin, the Calub gas field occurs (DU TOIT *et al.*, 1997; HUNEGNAW *et al.*, 1998).

4.- OLENEKIAN (245 - 243 Ma)

M. GAETANI[1]

with contributions of

V. LOZOWSKI, J. SZULC, A. ARCHE, F. CALVET & J. LOPEZ-GOMEZ

I.- MAIN FEATURES

This map (Fig. 4.1) has been added to the original project envisaged in the contract in order to respond to demands coming from authors of the adjacent maps for an improved supportive document of the facies and geodynamic evolution, not fully represented by the previous Wordian or successive Early Ladinian maps. This map only considers the northern Peri-Tethyan side where changes are more significant than in the south, which was largely emerged and under erosion. No deep Tethyan areas have been analysed.

Time slice definition

The Olenekian as stage of the Early Triassic was adopted by the Triassic Subcommission of the IUGS in its meeting of Lausanne in 1991. Since then, stage subdivision of the Triassic has been quite largely adopted, according to the following scheme (CGM, 1998).

	LATE	RHAETIAN NORIAN CARNIAN
TRIASSIC	MIDDLE	LADINIAN ANISIAN
	EARLY	OLENEKIAN INDUAN

Likewise, in the Tethys a substage nomenclature is also of common use. In the case of Olenekian, two substages have been defined, Spathian above and Smithian below. The map was drawn at the level of the Spathian, in correspondence of the *Subcolumbites* ammonoid zone and equivalent conodont *Neospathodus homeri* zone. These correlation tools can be applied from North Dobrogea to the central Caspian area, only. In the marginal or continental facies on the East European platform and other areas, the correlation was drawn using palynostratigraphy and vertebrates.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Two main kinematics trends are developing during the Olenekian:

- consumption of the Palaeo-Tethys along a northward subduction zone, have been active to the south of the present Caucasus and Kopet Dagh mountains. The Ural Orogenic Belt should have already been kinematically quiescent after the end of the convergence between Laurussia, Kazakhstania and Angara;

- dextral transcurrent regime between Laurussia and Gondwana, particularly well expressed along the margin of the East European platform;

The main continental blocks accounted for in this time are:

1.- the East European platform (EEP);

2.- the Pre-Ural and Pechora system of basins and depressions bordering to the east the EEP, filled by the clastics eroded on the Ural Range;

3.- the Pripyat, Dniepr, Donetz alignment of basins;

4.- the Precaspian basin and the Karpinky region, making its link to the Donbass;

5.- the system of troughs and basins that border the East European platform and the Baltic Shield from the North Atlantic rift to the Caucasus, via the Polish Trough and North Dobrogea;

6.- the collage of minor blocks, that would eventually consolidate to form the Turan plate with the Cimmerian orogeny;

7.- the microblocks of European affinity, like Moesia, Istanbul, Central and East Pontides, Dzhirula, that line the hanging wall of the subduction zone of Palaeo-Tethys;

8.- the Western Europe, in which an extensional regime with a significant N-S rifting is coupled with the pull-apart basins opening along the transtentional regime

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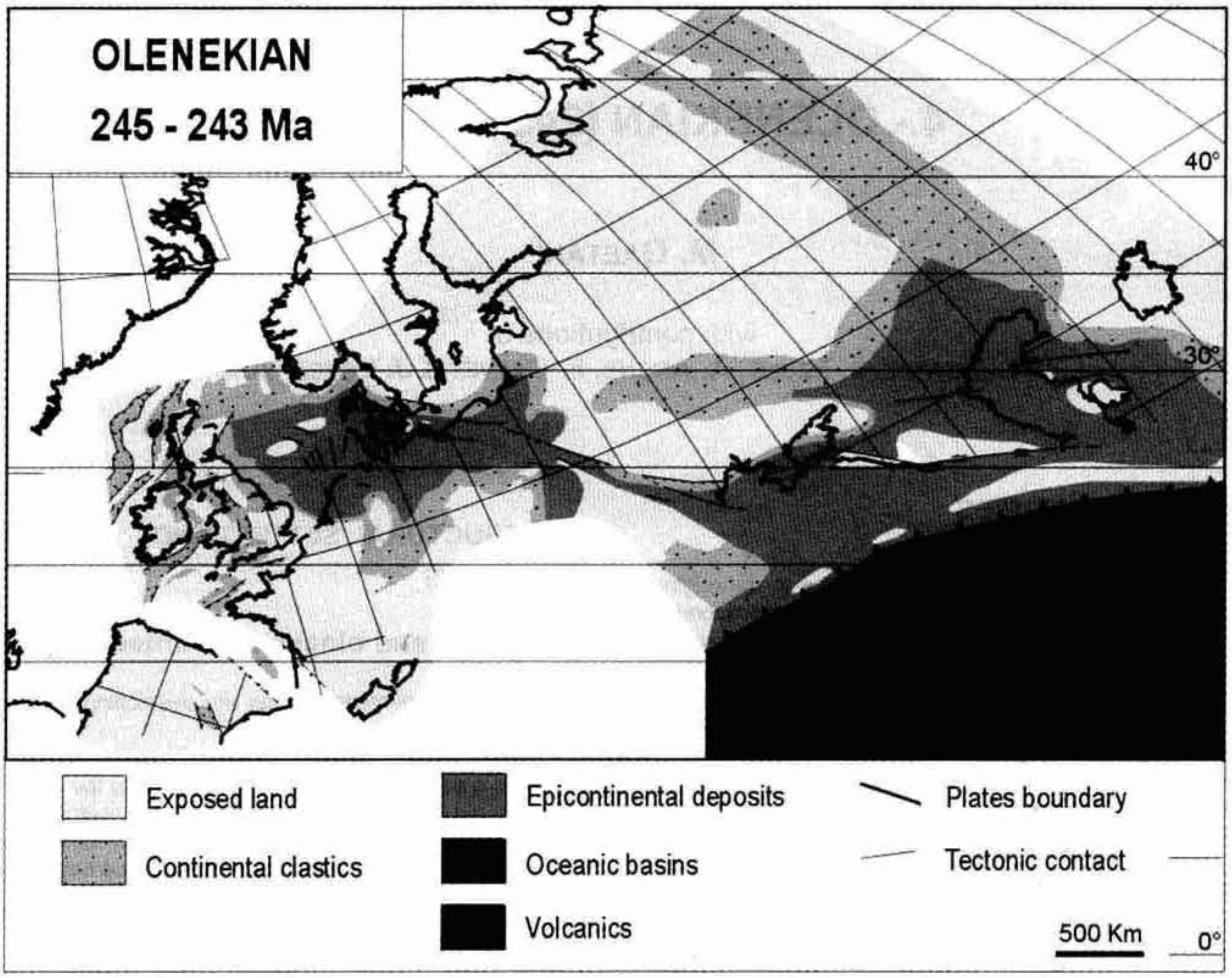

Fig. 4.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Olenekian.

active at the border of the Baltic Shield and East European platform;

9.- the British and Irish Block, that represent the southern extension of the north Atlantic rift system;

10.- the Iberia Block, separated by two transtentional alignments from Western Europe and the northern rim of Gondwana in Africa.

II.2.- Palaeoposition of plates and blocks

No new palaeolatitude data have been introduced in the present map. VAN DER VOO (1993), MUTTONI *et al.* (1996), TORCQ *et al.* (1997) and BESSE *et al.* (1998) discuss the data and Apparent Polar Wandering (APW) of Africa and Europe. Palaeolatitudes obtained by FEINBERG *et al.* (1996) on the Early Triassic (Induan, Dolnapa Svita) of Mangyshlak, Kazakhstan, have not been adopted in the present map. The calculated palaeolatitudes of 17° N ± 4 do not fit in with the general palaeogeographic picture obtained from the APW of major plates. There is a discrepancy of at least 10 degrees and the unrealistic model of LEMAIRE *et al.* (1998a and b) cannot account for such a bias.

II.3.- Accuracy

The inferred dextral transcurrent movement on the southern rim of the East European platform, was largely a dominant composer all along the parallels. No palaeomagnetic measurements might constrain such a movement. Missing magnetic anomalies, the accuracy of a palaeolatitude site is not better than 500 km. A number a small trenches and pull-apart basins that are supposed to be in existence during the transtensional regime, cannot be located with the palaeomagnetic method.

III.- DEFINITION AND DESCRIPTION OF THE DOMAINS

III.1.- East European platform (V. LOZOWSKI)

Moscow syneclise. The map is drawn at the level of the Yarenskian series, lower part, using as correlative tool the ganoid fish *Ceratodus* and the amphibian *Paratosuchus*. The continental succession consists of grey-green claystone containing lenses of cross-bedded

sand and sandstones. Characteristic of these series is the presence of siderite concretions. The maximum thickness is 40 m (LOZOWSKI, 1993).

III.2.- Pre-Ural and Pechora basins (V. LOZOWSKI)

III.2.1.- Southern part of the Pre-Ural

To the south, in the Belskaya depression, the Late Olenekian beds are defined as Petropavlosk Formation consisting of cyclic polymictic sandstones and minor amount of conglomerates with pebbles originating from the Ural mountains. Red shales and siltites overlay this first coarser horizon. Typical *Parotosuchus* fauna has been collected (SHISHKIN *et al.*, 1995). The thickness may reach a maximum of 250 m.

III.2.2.- Northern part of the Pre-Ural

In the Bolshesyninsk depression, the Byzovsk Formation consists of green-grey and yellowish cross-bedded sandstones with lens of conglomerates containing pebbles of magmatic and metamorphic rocks originating from the Urals. The maximum thickness is of 520 m. Here the correlation is also controlled by the *Paratosuchus* fauna (NOVIKOV, 1995).

III.2.3.- Pechora basin

The sediments are everywhere much finer than in the Pre-Ural basins and mostly consists of red-brown sometimes-grey shales, with intercalation of grey-green sandstones and siltites in the lower part. The depositional environment was largely in lacustrine facies (KALANTAR, 1987). The presence of siderite and limonite concretions is similar to the equivalent layers in the Moscow syneclise.

III.3.- Pripyat to Donetz basins (V. LOZOWSKI)

III.3.1.- Pripyat

Polychrome shales, light coloured marls and siltites with typical small ooidal concretions. The maximum thickness is 186 m in the upper member of the Mozyrsk Formation.

III.3.2.- Dniepr

Red shales and marls, rarely lacustrine limestone and kaolinitized sandstone. Maximum thickness 200 m in the upper member of the Radtchenkov Formation.

III.3.3.- Donetz

Red carbonatic shales rich in montmorillonite with frequent carbonate concretions, including a few cross-bedded polymictic sandstones. Upper member of the Adamovsk Formation with maximum thickness of 170 m (KISNERUS & SAIDAKOVSKY, 1972).

The correlations to the marine sections of the Precaspian basin are made of charophytes. There are reports of marine limestone in the folded Donbass Belt, that would suggest that temporary marine spell would occur also in the Donbass. However, they should be re-examined and confirmed.

III.4.- Precaspian basin (V. LOZOWSKI)

The wide Precaspian basin was the site of thick sedimentation during the Olenekian in a very smooth physiographic setting. On the external rim of the basin, continental, mostly alluvial but even lacustrine environments prevailed, while towards the centre of the basin, brackish and more open marine conditions gradually prevailed. Outcrops are present only on the external rim and exceptionally in the centre, like at Bolshoe Bogdo, a hillock not far from the Volga delta, due to the diapiric uplift of the Kungurian salt. From this locality originates the first Triassic ammonoid described in the Russian Empire (BUCH, 1831).

On the outer rim, the continental sediments are coarser towards the Ural Mountains, with cross-bedded fluviatile sandstones, characterised by the *Parotosuchus* fauna. Towards the west, far from the source areas, green shales and siltites dominates, with ferruginous and sideritic concretions. Towards the Volga river (Lipovsk Formation), shales and sandstone, occasionally marls and marly limestones, are reported. Thickness varies from 50 to 110 m.

Moving towards the centre of the basin and the Karpinsky area, influence of brackish environments became predominant and thickness dramatically increases, reaching usually more than 500 m only for the Late Olenekian. When reduced or missing, this is due to the halokynesis. Similarly, the position of environments did not remain stable because of significant lateral shifting. The transgressive tract of the sequence is defined by the *Dorikranites* ammonoid fauna, which represent the maximum extension of the marine facies. During the high standing tract, the basin was gradually filled and the marine environments gradually moved towards the deepest part of the basin, always represented by shales and siltites. The correlations are largely established on charophytes and ostracods. However, also ganoid fishes like *Ceratodus* and bones of amphibians washed from the subaerial flats, allow the establishment of links with the continental successions.

III.5.- Troughs and basins bordering the Baltic Shield and the EEP

III.5.1.- Polish Trough (J. SZULC)

The depocentre was aligned parallel to the Tornquist - Teisseyre lineament and was fed both from the Baltic Shield and by the Bohemian massif. Basically, there is an outer rim with continental deposits and a central basin with brackish sedimentation with short living more open marine spells. The rim towards the Baltic Shield held a wide embayment, the Lithuanian embayment, with coarser sandy alluvial sediments. The width of the alluvial rim and the size of clastics is reduced towards the East European platform, suggesting that the EEP was peneplaned at that time. In the Polish Trough itself, sediments were mostly reddish to green mudstones with sulphate crystals. The axial zone of the basin has been

divided into northern and southern segments by a system of swells (Sudetian - Malopolska - Lublin swells). In the southern subbasin, more open marine spells containing foraminifera and soft bottom bivalves such as *Gervillia* suggest a connection with the Tethys. Instead, the connections towards the north-west, inherited from the Zechstein basin, were gradually reduced. Correlations are largely based on palynomorphs that allow a fairly close and detailed stratigraphy (*Densoisporites neiburjii* and *Cycloverrutriletes presselensis* subzones). The careful quantitative analysis of the hygro vs. xerophilic palynomorphs (FIJALKOWSKA, 1994; FIJALKOWSKA-MADER, 1999) suggests that the climate was semi-arid with some short more humid intervals. The lithology supports this inference because of the reddish to green colour of sediments with caliche palaeosoils and sulphatic nodules.

III.5.2.- North Dobrogea

This area was tectonically very active during the Olenekian and several settings may be recognised. To the north of the Sfainctu Gheorghe Fault, evaporitic red-beds and shallow marine shales dominate, fed from the Ukrainian High. A deeper and particularly articulated basin is recorded in the Tulcea Unit (GRADINARU, 1995). The main facies are as follows:

- withish dolostones of shallow water.
- thin-bedded blackish marly limestone and marly shales with a *Tirolites* ammonoid fauna;
- allodapic limestones and graded beds of alternating fine-grained dark marly limestones and marls.

Slightly younger, but still belonging to the Late Olenekian, are red cherty limestones and red nodular limestone of the Hallstatt facies, rich in ammonoid, conodonts and ostracods (CRASQUIN-SOLEAU & GRADINARU, 1996).

The active faults also allowed the rising of basaltic magmas, to which are locally associated gabbroic bodies (lower part of the Niculitel Formation). The basalts show a MORB tholeiitic composition (SEGHEDI & OAIE, 1995; SEGHEDI, in press). Air fall pyroclastics and ignimbrites testify to volcanoes activity along the shoulders of grabens (SEGHEDI, 2000).

Along the eastern margin of the Macin zone, the presence of rhyolite necks as well as of vertical dolerite - rhyolite dyke swarms, suggests that here, too, volcanoes were active along the shoulders of the rifts (SEGHEDI, 2000).

III.5.3.- Crimea

Spotty evidence of marine shales and shallow water limestone is reported in boreholes from northern and central Crimea. No data on the southern Crimea is available.

III.5.4.- Scythian platform and Caucasus

The Olenekian rocks are spread and very significant in the NW Caucasus outcrops. The major tectonic slices contain diverse successions. To the south, at the headwaters of the Belaya river (Guseripl), the Olenekian sediments are missing and the area was under erosion. In the middle Belaya (Sakrai river) and in the Mali Laba basin, the Olenekian is mostly represented by thin bedded calcilutites (Yatyrgvart Formation) more than 200 m thick, containing in the upper part Late Olenekian conodonts (mostly *Neospathodus*, including *N. abruptus* and *N. triangularis*) (A. NICORA, pers. comm., 1999) as well as palynomorphs and spores of the Olenekian (YAROSHENKO, 1978). The Yatyrgvart Formation, and the subsequent other two formations of the Thach Group, the Maly Tchack and Achesbok formations, may be directly overlaid by a huge conglomeratic body, the Sakhrai Formation, considered as Late Anisian -Early Ladinian in age because of overlaying Middle Anisian carbonates with ammonoids (ROSTOVTSEV, 1973; OLEYNIKOV & ROSTOVTSEV, 1979; SHEVIREV, 1995). In addition to this evidence, on the left slopes of the hills along the Mala Laba river in front of the Nikitino village, M. GAETANI and E. GARZANTI found a conglomeratic horizon, up to 30 m thick, at 90% made of serpentinite pebbles, suggesting the rapid exhumation of a serpentinitic slab along transpressional faults during the Late Olenekian. This conglomerate is overlaid by oolitic limestone containing *Meandrospira cheni*, a foraminifer still of Late Olenekian age (R. RETTORI, pers. comm., 1999). To stress this point on the map, the area of the mid Belaya and Mali Laba is represented by a conglomerate fan, even if at least part of the time slice considered was under carbonate deposition.

To the north-east of the Stavropol High, an important basin developed on this part of the Scythian platform, known only through boreholes. The basin is subdivided in subbasins, the two most important being Kayazula and Mozdok (NIKISHIN *et al.*, 1998a and b). They are bounded by rifting margins and host marine sediments, mostly calcareous with shales intercalations to the north and more terrigenous to the south, with some volcanoclastic influx. Single isolated amounts of intraplate basalts linked to the rifting have been also reported. The general geodynamic setting seems to belong to an extensional regime linked to the back arc conditions. The dextral transcurrent movements could be responsible for the *en échelon* distribution of the basins and uplifts. Scanty data are available on the possible occurrence of Olenekian rocks in the Kuban basin, around Maikop.

III.6.- The Turan area

The area that will be consolidated in the Turan Block in the Late Triassic, experienced during the Olenekian different geodynamic settings. Some parts like the North Ustyurt and the Karabogaz were emergent and, in the case of the Karabogaz, deeply dissected down to the crystalline pre-Ordovician basement. Others, like the Busachi peninsula or the Karakum, were sites of low subsidence with a veneer of shaly sediments deposed under alluvial or marginal facies (THOMAS *et al.*, 1999). In the Gornyi and South Mangyshlak instead, strong subsidence was acting, the Late Olenekian being represented by a non-decompacted thickness of sediments not less of 750 m. Since sea depth probably never exceeded 100 m, and considering the Olenekian not longer than 3 Ma, and the Late Olenekian duration conventionally of only 1.5 Ma (assumption based on the following values: 250 ± 1 Ma for the P/T boundary and 241 ± 1 Ma for the base of

Ladinian), the accommodation rate is around 500 m /Ma. Such huge subsidence is thought to be controlled by the extensional/transtentional regime linked to the back arc setting. Sedimentation was mostly shaly and silty, red-brown in the lower part and the upper parts, dark grey in the middle. Carbonate beds are rare in the lower part, with typical shelly tempestitic layers at the very base of the Tyururpa Group, containing also the ammonoid *Dorikranites,* which is the best correlation tool in the area. In the middle part, carbonates form concretions locally filled up with ammonoids (SHEVIREV, 1990; BALINI *et al.*, 2000). The Mangyshlak formed an embayment open to the west while passing to more proximal conditions to the east (SCHLEZINGER, 1965; LIPATOVA, 1984; ZHIDOVINOV, 1994). The carbonate intercalations are scanty and limited to the basal *Dorikranites* layers, that may be interpreted as the transgressive tract of the sequence, and to the basal part of the Columbites Beds, that might be considered as the high standing tract.(GAETANI *et al.*, 1998). The rare fine sandstones are volcanoclastic, largely derived from the erosion of intermediate to felsic volcanic products (dacitic). Their position is supposed to lie not very far, on the northern flank of the Karabogaz high, connected to the extensional faulting along the Mangyshlak trench. This back-arc basin is proved to continue in the Tuarkyr area, for the finding of *Dorikranites* fauna (LUPPOV, 1957; GARZANTI & GAETANI, 2000).

The main volcanic arc was situated to the south, not far from the region of Aghdarband in NE Iran (RUTTNER *et al.*, 1991; BAUD *et al.*, 1989, 1991). Here the Spathian section consists of a basal tuffitic breccia intercalated with brecciated lava flows, overlaid by 150-200 m of shallow water massive limestone (Sefid Kuh Limestone) with rare conodonts. Here too, the very high sedimentation rate of carbonates is linked to the back-arc setting.

III.7.- Microblocks and Küre basin

No detailed information are available on the wide marine seaways that it is interpreted to lie to the south of the transtentional belt. The only possible evidence is provided by the Dizi series that could also contain rocks of Olenekian age in its turbiditic sequences. However, it seems that up to now only the Late Triassic has been proved by spores (SOMIN & BELOV, 1967), even if these findings are discussed by OLEYNIKOV & ROSTOVTSEV (1979).

Both on Dzhirula and in Central and Eastern Pontides, continental microblocks do not have evidence of Olenekian sediments and are considered as emergent. In the Istanbul zone part of the area was still emergent, but in the Kokaeli peninsula alluvional succession with conglomerates and coarse red cross-bedded sandstones is overlaid by Early Anisian nodular limestone rich in ammonoids (stratotype of the substage Bithynian) (ASSERETO, 1974). Hence, by geometric position, it is possible that at least in part these conglomerates are of Olenekian age. However, according to S. DERMAN (pers. comm., 2000) the Tepekli conglomerate is separated by the Anisian nodular limestone by an angular unconformity.

Moesia was largely interested by alluvial sedimentation. Coarse polymictic conglomerates and sandstones with red matrix are cropping out all along the Fore Balkan and Balkan Belt. Alluvial sandstones grading up in shales, instead, have been drilled on the Moesia block itself (TARI *et al.*, 1997). This succession in the Bulgarian east part of the Moesia platform gradually grades to evaporitic beds. However, age definition is poor. Apparently, a roughly E-W trend could be observed, with the sea prograding from the east.

III.8.- Western Europe

The map was draft above the Hardegsen unconformity, in the Solling Formation and equivalents, forming the lower part of the Solling sequence, still considered as Middle Buntsandstein by AIGNER & BACHMANN (1992) or already Late Buntsandstein by GELUK & RÖHLING (1997, 1999) and GELUK (1999). Data are very tightly controlled through a number of released boreholes and very detailed isopach and facies maps that have been recently issued in the quoted papers. In the basin the sequence has developed entirely in a clay- to siltstone facies, whist on the border coarser, fully continental facies developed. Coarser fractions are known nearby the Bohemian massif and the Baltic Shield, whilst much finer sediments originated from the Central and the Armorican massifs. Typical of the Mid European basin is the presence of grabens with rift margins oriented N-S, which rotate to the WNW-ESE direction when approaching the Polish Trough and the belt still affected by minor dextral transcurrent movements along the margin of the EEP. The regional subsidence along the Tornquist zone in Denmark was gently dipping towards SW, whilst minor faulting and differential subsidence occurred associated with the Boerglund Fault (MOGENSEN, 1995). Prior to the Solling and equivalent units, several pulse of tectonic activity produced the Hardegsen unconformity, which is known through most of the area.

To be also noted that during the Olenekian there were no connections towards the North Atlantic and towards the south through the future Burgundian gate. The most obvious connections to marine seaways were towards east, through the Moravian gate and perhaps, but evidence is poor, through the East Carpathian gate.

III.9.- British and Irish Isles

This region forms the west boundary to the mid European basin and its western side. It is interested by a series of grabens oriented NE-SW, which are aligned with the North Atlantic system of rifts. Sedimentation was invariably in continental alluvial settings and grouped within the Sherwood Group, oftenly displaying coarse sandstone and microconglomerate size (WARRINGTON & IVIMEY-COOK, 1992). Main depocentres were to the north of the Welsh and London - Brabant massifs.

III.10.- Iberia (A. Arche, F. Calvet & J. Lopez-Gomez)

The precise identification of the Olenekian in Iberia is not easy, because most of the age assignments are based on facies resemblance to the Mid European Buntsandstein or on sequence stratigraphy interpretation. The main outcrops are found in the Iberian basin within

the Iberian ranges. Minor outcrops appeared in the Murcia area and in the Ebro coastal area. A number of depositional sheets have been identified (Canizar Formation) (LOPEZ-GOMEZ & ARCHE, 1993) in the arkosic sandstones of the fluvial sheets that dominated the sedimentation. Upward in the succession, this sequence grades into finer sediments and red mudstones are interbedded within the sandstones, the latter interpreted as remnants of point bars and traverse sand bodies (Eslida Formation). Thickness may be fairly significant with several hundred of metres. The difficulty to have precise dating hampers better correlations.

5.- EARLY LADINIAN (238 - 235 Ma)

M. GAETANI[1]

with contributions of

V. LOZOWSKI, J. SZULC, A. ARCHE, F. CALVET, J. LOPEZ-GOMEZ & F. HIRSCH

I.- MAIN FEATURES

The standard subdivisions of the Triassic are of almost world-wide usage, especially as far as the Middle Triassic stages, i.e. Anisian and Ladinian stages, are concerned. Still unsettled is the formal definition of the very base of the Ladinian. In the present map, the Early Ladinian is considered as corresponding to the *Eoprotrachyceras curionii* zone in terms of Tethyan ammonoid zones and its equivalent *subasperum* zone in the Boreal domain. If in the Tethys this time slice often allows fine correlations, in the marginal areas where ammonoids and conodonts are scanty or absent, such correlations are not always very precise. Magnetostratigraphy is now well defined in the Tethys (MUTTONI *et al.*, 1998). However, at present, no results have been obtained in the continental and marginal Peri-Tethyan areas like in the Germanic basin, where the attempts made have seemingly been successful. Palynostratigraphy is also of use and the *plurianulatus* - *secatus* phase is considered as correlative for the Early Ladinian *Curionii* zone (BRUGMAN, 1986).

More than a century of researches allows to have a sufficiently firm correlation between the Mid European basin (here used for Germanic basin) and the Tethys. Our map is drawn in the Mid European basin at the time corresponding to the earliest part of the evolutionary lineage of the genus *Ceratites*, in the Upper Muschelkak, lower part of the *Ceratites* Schichten (HAGDORN, 1991). This part of the Upper Muschelkak corresponds also to a sea level high standing, in terms of sequence stratigraphy (AIGNER & BACHMANN, 1992). The correlations of the map are fairly accurate in the Mid European basin, but their accuracy decrease dramatically eastwards. Only in North Dobrogea it is still reliable at the chronozone level. The accuracy is often bounded to the stage level in the continental facies bordering to the south the East European platform and in the Precaspian basin. On the southern shores of Tethys, large emersions and continental or very restricted evaporitic facies make often merely tentative to establish the age of the sedimentary successions. Only in Israel, correlations are fine (ZAK, 1986).

The corresponding Tethys map (MARCOUX *et al.*, 1993a) was drawn for the Late Anisian. In this project the selected time slice was shifted to the younger Early Ladinian, because of the wider marine ingressions on the marginal areas, due to the high standing which seems to characterise the Early Ladinian. The differences in terms of Tethys evolution are not relevant. If we accept the geochronology of MUNDIL *et al.* (1996), based on U-Pb dating on zircons, the difference of time is only of a couple of Ma.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Three main kinematic trends might be outlined during the Ladinian:

1.- opening of the Neo-Tethys, with the anticlockwise rotation of the Cimmerian blocks and the their quick displacement towards the Eurasian active margin, achieving for the Iranian Spur a pre-collision condition (BESSE *et al.*, 1998);

2.- the dextral transcurrent regime between Laurussia and Gondwana probably still existing, but significantly lowered down in comparison to the Permian times, and

3.- the opening of Neo-Tethys generated also a westwards propagation of the Tethyan branches, further dissecting the margin of Europe as a result of the Hercynian orogenesis.

An additional point is linked to the general marine aggradations observed on the northern shore of Tethys, whilst on African and Arabian margins, the trend seems instead less homogeneous. The open marine facies are more reduced than in the Late Permian on the Arabian Peninsula. The high standing of the Early Ladinian is well expressed in Israel (Ramonin Formation), while westwards, sea aggradation was very limited. Since the world-wide trend in the Early Ladinian is for high standing, this

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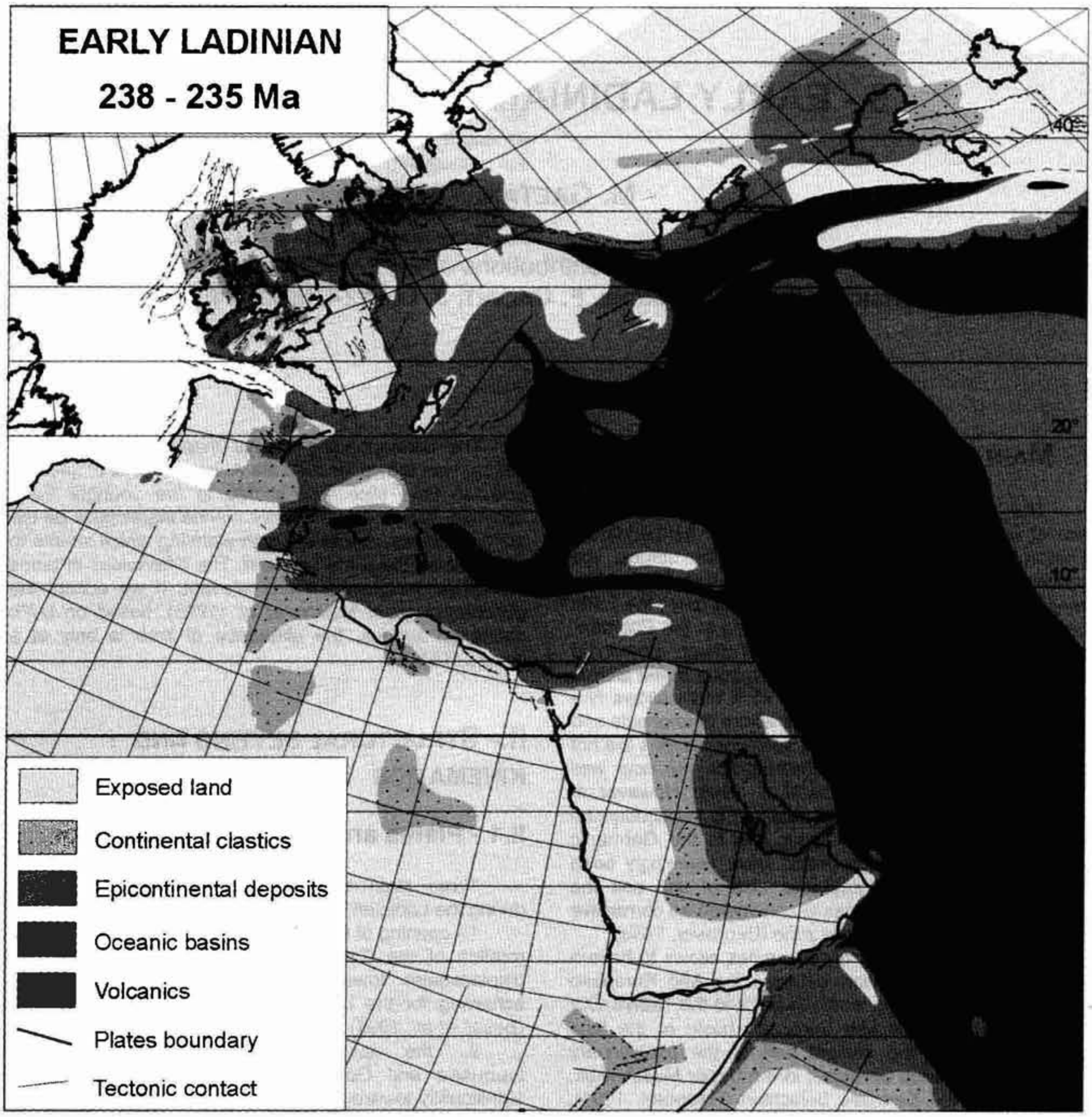

Fig. 5.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Ladinian.

African and Arabian regression should be linked to a positive movement and also perhaps to a non-sufficiently accurate dating.

The main continental blocks accounted for in this time are:

1.- the East European platform (EEP) in which the Dniepr - Donetz alignment of basins experience a significant lowering, even a stop, of its subsidence, before the Late Triassic inversion. Possibly, the Karpinsky depression already starts its inversion in the Ladinian;

2.- the system of troughs and basins that border the East European Platform and the Baltic Shield from the North Atlantic rift to the Caucasus, via the Polish Trough and North Dobrogea. From North Dobrogea to Caucasus it evolved in a back-arc basin, behind the subduction front of the Palaeo-Tethys;

3.- the collage of minor blocks, that will eventually result in the Turan plate after the Cimmerian orogeny;

4.- the microblocks of European affinity, like Moesia, Istanbul, East Pontides, Dzhirula that line the hanging wall of the subduction zone of Palaeo-Tethys;

5.- the Western Europe, in which an extensional regime with a significant N-S rifting is coupled with the pull-apart basins opening along the transtentional regime active at the border of the Baltic Shield and East European platform. However, this transtentional regime is gradually decreasing from the Olenekian to the Ladinian in its importance;

6.- the Iberia Block, separated by two transtentional alignments from Western Europe and the northern rim of Gondwana in Africa;

7.- the Gondwana margin, mostly with a passive continental margin behaviour.

II.2.- Palaeoposition of plates and blocks

The only new datum concerning palaeoposition computed with palaeomagnetics originates from the paper of MUTTONI *et al.* (2000), concerning the Fore-Balkan in the locality of Belogradchik (Bulgaria). Being the Fore-Balkan a gently deformed range proxy to Moesian Block, we assumed that the palaeolatitude of the Late Anisian rocks of western Fore-Balkans, comprised between 21° and 24° N, could be the palaeolatitude of the Moesian southern margin. This figure supports the interpretation adopted to position the Moesian Block to the SE of the present position, widening the North Dobrogea transtensional and back-arc seaway.

No available new additional data exists for the APW of major blocks.

II.3.- Accuracy

The accuracy of correlation on the biostratigraphic base has been already discussed. As physical stratigraphy is concerned, magnetostratigraphy is at present available only inside Tethys and not on its bordering shelves. The poor dating of many successions hamper correlations based on sequence stratigraphy.

The data on the Maghreb should be probably referred to a slightly younger time. We introduced them in the map to show the progressive aggradation of the depositional sequences on the previously emerged margin.

III.- DEFINITION AND DESCRIPTION OF THE DOMAINS

III.1.- East European platform (V. LOZOWSKI & M. GAETANI)

Of the 3 sub-basins active during the Permian, only the Precaspian basin was still actively subsiding.

The Pripyat basin had local residual infilling of red and green clays that may be attributed to the Ladinian. The Dniepr basin is considered already inverted or at least filled up and only the Donetz basin was still possibly accommodating some alluvial deposits (KISNERUS & SAIDAKOVSKY, 1972). To the east of the present 38° E of longitude, Triassic strata are eroded away in the Donbas fold belt and most probably the same holds true for the Karpinsky former depression (STOBVA & STEPHENSON, 1999). Indirect evidence of the beginning of the inversion is represented by the increasing in the clastic input to the nearby border of the Precaspian basin (ZHIDOVINOV, pers. comm., Moscow, 1998).

The Pre-Uralian depression was also almost filled up during the Middle Triassic and only a narrow strip of fine clastic sediments may be traced between the front of the Ural Range and the emerged East European Platform.

The connection between the southernmost part of Pre-Ural depression and the north-east tip of the Precaspian basin is still occupied by terrigenous successions of sandstone and shales, mostly grey in colour.

The wide Precaspian basin was a large shallow water depression linked to the open sea of Tethys through the gate between the Ukrainian High and the Stavropol High. Its size is comparable with the Middle European basin. Ladinian rocks are known only through boreholes, but a small part of the south-east margin in Mangyshlak. The facies show a bull-eye distribution, with an external rim of alluvial plain, followed by brackish water characterised by the ostracod *Gemmanella* and with a central part with normal salinity. Salinity level may change fairly largely, and this explains the rather different facies interpretations according also to the accuracy of time-equivalences (LIPATOVA, 1984; ZHIDOVINOV, 1994; KUKHTINOV & CRASQUIN-SOLEAU, 1999). The correlations within the basin are largely based on ostracods, charophytes, and bivalves thriving on soft bottoms. Sediments are grey calcarenitic and marly limestones, siltites and shales. The terrigenous influx increase towards the upper part of the Early Ladinian, as an answer to inversion activity that is going on at the south of the basin along the alignment Karpinsky - Mangyshlak. Total thickness is usually of few hundred meters, though the control by the halokynesis of the Permian salt may be locally significant.

III.2.- Border basins at the margin of the EEP

III.2.1.- Scythian platform and Caucasus

The Scythian platform data come only from boreholes. They have been recently summarised by NIKISHIN *et al.* (1998a and b). West to east, three main areas may be distinguished: the central-northern Crimea, the Kuban basin in the west Pre-Caucasus and the east Pre-Caucasus basin, subdivided in several sub-basins.

According to NIKISHIN *et al.*, these basins are driven by a basically orthogonal rifting mechanism, but the further interpretation of the Caucasus outcrops suggests that also a transtensional component should be present in the activation of these depocentres (GAETANI & GARZANTI, in progress).

In the Simferopol area (Crimea), shallow water limestone are tentatively attributed to the Ladinian, and a similar facies is also supposed for the west Pre-Caucasus, but no firm data are available. In the complex of sub-basins in the eastern Pre-Caucasus, marls, siltites and shales filled up the more northern sub-basin (Kayasula), whilst in the south (Mozdok), also carbonate rocks are interbedded. Along the faults that bounded the Kayasula basin, isolated basalts poured out.

The Caucasus continued to be interested by significant tectonic activity. In the NW, the areas of the rivers Belaya and Laba were sites of emersion and

erosion (Guseripl) or deposition of conglomerates and sandstones, mostly fed by the metamorphic basement, with a residual serpentinoschist component. According to SHEVIREV (1995), who claims for the necessity of a revision of the Middle Triassic stratigraphy in NW Caucasus, Ladinian rocks might be in places even totally absent. In Mala Laba area, Late Triassic rocks of the Podkhodz Group may lie unconformably on Early Triassic Maly Thach Formation (YAROSHENKO, 1978). As a matter of fact, there is no palaeontological evidence for the Early Ladinian. Only Late Ladinian ammonoids have been reported in rocks overlaying the main terrigenous body.

III.2.2.- Polish Trough (J. SZULC)

The region between the EEP and the Bohemian massif was characterised by a fairly wide homoclinal slope to the north, dipping towards the two aligned depocentres of the Mid-Polish Trough and the Foresudetic Depression, actively controlled by the Odra Fault (SZULC, 1993). Minor local depocentres aligned along the Tornquist - Teisseyre Fault system are linked to the halokynetics of the Late Permian Zechstein, forced by the strike-slip motion along the fault. The deepest, open marine sedimentation occurred within the Foresudetic depression, which constitutes the easternmost part in Poland of the Middle European basin. It worked out a link through the Silesian Gate, whilst the Moravian Gate closed during the Late Anisian. To the east, a connection with the Dobrogea - Küre - Dizi back-arc sea is interpreted to exist through the East Carpathian Gate.

The deepest, open marine carbonate sedimentation took place within the intensively subsiding Foresudetic Depression. The fauna in the micritic limestones of this subbasin is typical for its high density and low diversity, suggesting somewhat restricted conditions. On the shallower ramp areas, marly sedimentation rich in glaucony prevailed. The onshore belt displays increasing clastic feeding. Accuracy of correlation is fairly good, at least for the more open marine areas (upper Silesia and Holy Cross Mountains), where gondolellid conodonts and ammonoid *Ceratites* spp. were found. To be noted, the absence of red beds and the rare salt deposits, in contrast to the western part of the Middle European basin. This fact could be linked to local more humid climate or/and to fresh water influx from the nearby EEP.

III.3.- Turan area

Most of the area is now emerged or supposed to be. Evidence is often missing because of latest Triassic and Early Jurassic erosion. North Ustyurt, Karakum and Karabogaz are thought to be emergent as well as the smaller island of Buzachi. However, deposition of Middle Triassic sediments and later erosion cannot be excluded. The sedimentation was mostly concentrated in Mangyshlak and its neighbouring areas, where continental conditions prevailed. In Gornyi Mangyshlak, the Ladinian is represented by alluvial plain cycles, with fine and medium sized volcanoclastic sandstones forming the point-bar of the meander systems (GAETANI *et al.*, 1998). The huge amount of volcanoclastics with mostly dacitic signature is peculiar to this area and considered to be fed by a not very far source, also because there is a minimal contamination by a basement clastic source. The position of these volcanoes is supposed to lie between the Karabogaz High and the Mangyshlak Depression, because otherwise contamination from the basement should be revealed in the clastic infilling of Mangyshlak. The back-arc setting, interpreted to be active in the area from the Early Triassic to the earliest Norian (GAETANI *et al.,* 1998; THOMAS *et al.*, 1999), controls the subsidence accommodating several hundred meters of alluvial sediments.

III.4.- The North Dobrogea - Küre - Dizi back-arc sea

Along with the ongoing subduction of the Palaeo-Tethys, a back arc basin originated between the margin of the East European Platform and the subduction-related arc. This elongated back arc was extended from North Dobrogea to the Transcaucasus. Their sediments are poorly preserved only as small remnants in Dizi Flysch, south of the Caucasus Range, in the Tauric Flysch of Crimea, in the fragments of Tauric Flysch, possibly present in the Pontides, and in the more articulated mostly deep sedimentation of North Dobrogea. The last is the sole area where sediments are sufficiently well preserved to allow a more sounded interpretation.

III.4.1.- Dizi Flysch

The upper part of the Dizi series consists of thin bedded, well sorted, fine grained sandstones and shales, often including tuffaceous material (SOMIN & BELOV,1967; KAZMIN, 1990). Direct evidence of Ladinian age is however missing.

III.4.2.- Küre basin

The present map adopted the interpretation of USTAOMER & ROBERSTON (1997, fig. 2) and, consequently, the Küre Ophiolite is considered as a remnant of the back arc basin to the north of the Palaeo-Tethys subduction zone. The Küre Ophiolite is thought to be covered by Permian - Early Jurassic sediments, but no Ladinian sediments are known from this unit (USTAOMER & ROBERSTON, 1997).

III.4.3.- Gornyi Crimea

Exotic block embedded in Tauric Flysch have been known for a long time in Southern Crimea. The unsettled problem is to establish for sure the age of the base of Tauric Flysch. Even the recent paper by KOTLYAR *et al.* (1999) leaves open the question as whether the assemblage and locality with Anisian brachiopods described by DAGYS & SHVANOV (1965) is a block or represents an episode of Anisian sedimentation of the Tauric Flysch. In the event that this second interpretation be accepted, a Middle Triassic age for the onsetting of the Triassic flysch would be confirmed.

III.4.4.- North Dobrogea

Stratigraphy and facies pattern have been recently summarised by GRADINARU (1995).

The North Dobrogea shows in a now telescoped nappe structure a transect from shallow water carbonates to deep-water pelagites. From south-west to the north-east in the present orientation, the facies are the following. In the Macin nappe, thin bedded, grey micritic limestones and whitish bioclastic limestones are represented. In the adjacent Niculitel Unit most interesting is the occurrence of a huge pile of pillow basalts, geochemically considered as the product of an intraplate volcanism (SAVU, 1986). The age of these Niculitel basalts is critical. According to GRADINARU (1995) and SEGHEDI (2000), the age is restricted to the latest Early Triassic and to the Anisian. However, the Cataloi Formation overlying unit, is reported to be Late Ladinian to Norian and the siliciclastics of the Alba Formation are reported to start with the Late Carnian (SEGHEDI, 2000). The basalt outpouring is here considered as linked to the thinning for the extension of the back arc (SANDULESCU, pers. comm., 1998). The area has been interpreted in the present map as oceanic-like crust, still under an extensional setting. However, direct evidence for rocks for Early Ladinian age in the Niculitel Unit is missing. In the Tulcea Unit, where the best development is exposed, the Early Ladinian is represented by red nodular limestone, red cherty nodular limestone and shale deposed in a pelagic environment with low to very low sedimentation rate. The Agighiol ammonoid fauna represents a classic for the Hallstatt-type facies since the beginning of the century (KITTL, 1908; SIMIONESCU, 1914).

III.5.- Microblocks of European affinity

1.- Dzhirula and Khrami in the Transcaucasus, slabs strongly deformed during the Hercynian Orogeny with evidence of intrusive and sedimentary rocks of Palaeozoic age, probably emergent during the Ladinian (ADAMIA & LODKIPANIDZE, 1989).

2.- The Eastern Pontides contains a metabasite - phyllite - marble unit, tentatively considered of Permo-Triassic age (OKAY & SAHINTÜRK, 1997).

3.- In the Western Pontides, where the Istanbul Zone or Unit is preserved (OKAY, 1989; OKAY & SAHINTÜRK, 1997), the Early Ladinian is represented by grey to pink-red nodular limestone, well exposed in the Kokaeli peninsula (ASSERETO, 1974). In the Central Pontides, the Devrekani Metamorphic Unit might represent the continental crust core of the part of the arc to the south of which Palaeo-Tethys was subducting (USTAOMER & ROBERSTON, 1997).

4.- Chios. In the present map, the Chios island is considered to be proxy to the arc since the late Early Triassic. Of the two major nappes in Chios, the so called allochthonous near Marmaro, a "terra rossa" horizon, lies between the shallow water Permian carbonates and the overlying platform carbonates of Jurassic age. In the so-called "autochthonous" shallow water carbonates are recognised, as well as deeper shaly and cherty rocks (GAETANI *et al.*, 1992; MUTTONI et al., 1994).

5.- Moesia. The Moesia block and its proxy Fore-Balkan were largely submerged by a shallow water carbonate ramp open to the north towards the Dobrogea basin. In the north of Bulgaria and in Romania this is represented by limestones and dolomitic limestones, whilst to the south, mostly under the Bulgarian plain, a more restricted facies prevail. Here, a more consistent silty/shaly influx is characteristic (GEORGIEV & ATANASOV, 1993; TARI *et al.*, 1997). To the east in Bulgaria, huge anhydrite and salt bodies are reported, mostly formed during the Carnian (GEORGIEV, 1996). The evidence for an earlier development in the Ladinian is scanty. The North Bulgarian uplift and its prosecution in Romania, in the area of Constantsa, represent a significant physiographic feature. This uplift might be considered as the shoulder of an asymmetric rifting aligned E-W in present position, possibly starting in the Ladinian, but mostly developed in the Carnian (GEORGIEV, 1996; TARI *et al.*, 1997). To the north, the transition to the Dobrogea deep basin is only interpretative, due to the nappe structure of North Dobrogea. To the west and to the south, a gradually deepening margin is hypothesised towards the Tethys. The only direct evidence consists in the open platform carbonates of the western Fore-Balkan and the Balkan to the south.

III.6.- Middle European basin

The Polish extension has been already described. The bulk of the basin is located in Germany where the Upper Muschelkalk represented it. It consists of numerous shallowing-upwards m-thick cycles (AIGNER & BACHMANN, 1992). They have a shaly base, overlaid by calcareous tempestites and a calcareous top with bioclasts, ooids and intraclasts. These parasequences are organised in wider sequence system tracts, of which the so-called "Cycloides-bank" (a terebratulid brachiopod) could represent the starvation associated to the maximum flooding surface. Moving to the margins of the basin, shaly shore face sediments occur along the northern rim, whilst the more carbonatic, locally with evaporite interbedding like in Luxembourg or with brackish water clays, are more frequent to the south, in the border of the Burgundian Gate. The gate was a more efficient connection with the Tethys during the Ladinian, in comparison to the eastern Carpathian Gate, whilst the Moravian Gate was no more in function since the Late Anisian (SZULC, 1999: BRACK *et al.*, 1999).

A typical marginal succession may be studied in Ardèche at the Largentière/Balazuc area, where intensive mining and prospecting studies ware carried out (COUREL, 1984).

III.7.- British and Irish Isles

The Mid European basin carbonatic facies gradually passes to a wide shallow water muddy depositional system (Mercia Mudstone Group, WARRINGTON & IVIMEY-COOK, 1992). Local lateral intercalations of mudstone sulphates and halite, which are interspersed within the Mercia Group, are probably slightly older than the Early Ladinian. Notwithstanding, biostratigraphic control, mostly based on miospores, is poor. Landwards, mature meandering fluviatile systems were still active, especially to the south, separating the Welsh massif from the London - Brabant massif. Together with the Northern England, these emergent areas formed the divide between the Mid European basin and the complex of

grabens with shallow marine muddy facies, laterally passing to brackish water, that characterise the depressions aligned along the North Atlantic system of rifting. A connection with the open sea is supposed towards the Celtic Sea at SW.

The map shows a number of rifting structures in this area as well as in the North Sea and Mid European basin. The very dense covering of seismic exploration has shown a number of faults linked to the regional tensional crustal stresses. However, the actual dating of the tectonic lineament is often hampered by the poor biostratigraphic control. Therefore, in agreement with WARRINGTON & IVIMEY-COOK (1992), the structures may have been active in Triassic, with no implicit movements at one particular time or throughout the period.

III.8.- Iberia (A. ARCHE, F. CALVET, J. LOPEZ-GOMEZ)

In Iberia the facies distribution is rather simple. North and west emergent areas feed the depositional area in the east. The Molina de Aragon corridor is possibly fed by a north-west source, raising the question of the significance of the Bay of Biscay Rift. For the alluvial deposits in the Catalan domain, the source is located to the NE, from Palaeozoic outcrops in the eastern Pyrénées. Two other alluvial plains are located at the margin of the Iberian massif. These continental deposits grade into peritidal silicoclastic tidal flats, deltas and evaporitic sebkas, and eventually into shallow, epicontinental carbonatic ramps to the east (LOPEZ-GOMEZ & ARCHE, 1993; CALVET *et al.*, 1990; LOPEZ-GOMEZ *et al.*, 1993).

The tectonic regime for the Early Ladinian was one of early thermal subsidence, except for the Catalan domain in the NE, where rifting was still going on and even limited volcanic activity briefly took part in its southern part.

III.9.- Gondwanan margin

III.9.1.- Maghreb

A vast Peri-Tethyan domain covered the NW Africa during the Triassic, with the sea onlapping from the north-east and north during the Middle Triassic. However, the boundary of the Peri-Tethyan domain to the north is unknown, being covered all along the present coastal area by the overthrusted allochthonous Maghrebides. Being the amplitude of the displacement difficult to evaluate, the junction between the epicratonic Peri-Tethyan and the margin of the Tethys seaway is somehow problematic. To the south instead, towards the cratonic domain, the sedimentation area and the depositional settings are clearly recognisable. Due to the marginal facies, the age assignment may be largely non accurate. Clear evidence is only available for the Late Ladinian. In the present map, the facies distribution is that of the Late Ladinian thought also to have possibly started earlier in Tunisia and in the Oujda area in Morocco. The Late Ladinian ingression ends westwards roughly along the present 8° E longitude. In the Tunisia, the South Dahar High divides the brackish to evaporitic flat to the north, from the mostly continental fluviatile and fluvio-lacustrine flat in the nearby Libya. To the north, the borehole of Rheouis revealed shaly/anhydritic levels, displaying also clear carbonate marine deposits (COUREL *et al.*, 2000). On the contrary, dolostones were drilled at the Cap Bon, near Tunis.

In Morocco, the graben of Oujda contains the oldest Triassic rocks of the area, with carbonates associated with basalt flows (OUJIDI, 1994). Volcanic activity is supposed to have existed also in Tunisia, west of Cap Bon and to the south-west of the South Dahar High.

III.9.2.- Libya and Egypt

The region may subdivided in two parts, separated by an alignment of positive structures, like the Gargaf High, which remained under the erosional setting. To the south, two major internal basins, the Murzuk and Al Kufrah basins, contains the so-called Post-Tassilian Nubian Sandstones, which may span from the Late Permian to earliest Cretaceous. In the Al Kufrah basin, Anisian sporomorphs have been detected, but no Ladinian (BELLINI *et al.*, 1991). In the map, the presence of Ladinian sediments is only tentative since persistence in the deposition has been assumed.

To the north, the situation is more articulated. In Tripolitania, a wide continental fluviatile and fluvio-lacustrine flat extends from the Zarzaitine ridge (GRUBIC *et al.*, 1991), in Algeria near the boundary with Libya, to the Tripolitanian Gefara and to the Homra basin. Here continental, brackish and evaporitic conditions are interfingering. Ladinian sediments are preserved in the Ras Hamia Formation with dolomitic shales, marls, and dolostones with brackish water bivalves (MENNING *et al.*, 1963). Similar shales and siltstone are exposed in the Currusc Formation in the Gefara (ASSERETO & BENELLI, 1971). The marginal facies extends to the south of Tunisia in the Djefara area, whilst the southernmost tip of Tunisia was covered by the fluviatile apron, with fine sandstones, like in the area of the El Borma oil field.

The Syrte basin, which developed mostly later in the Cretaceous, was interested by a rifting originating a NW-SE / W-NW - E-SE oriented depression filled by clastic continental sediments, of which a maximum thickness of 270 m is attributed to the Ladinian. A wide flat is developed and the transition to marginal and shore face condition is suggested towards the NW of the basin. Transition from coarse sandstone and fine conglomerates to fine sandstones and shales, with dolostone intercalations has been observed (BELHAJ, 1996). Pollen grains, as well as a few acritarchs have been detected (BRUGMAN & VISSHER, 1988). The Cyrenaica acted as a rift shoulder both at the north-east of the Syrte basin, and in another extensional structures towards the present Mediterranean, according to seismic interpretation and few boreholes that reached such depths (DEL BEN & FINETTI, 1991).

The Western Desert area of Egypt was mostly under erosional conditions and the Delta area was never penetrated below the Cretaceous. In the Matruh Basin and Gebel Rissu basin, arenitic deposits are tentatively considered of Middle Triassic age (KEELEY, 1994). GUIRAUD *et al.* (2000) suppose the existence of a marginal sea already below the Western Desert, but data are not

available. It is also difficult to assess if the tectonic rifting that could be detected during the Jurassic, might be considered as already active during the Triassic, owing to missing sediments. Consequently, the palaeogeographic and tectonic reconstructions for the southern margin of the present Eastern Mediterranean are mostly dependent upon the adopted model of geodynamic evolution. In the Sinai and Wadi Araba the margin towards the Levant depositional basin is present, as described in the Levant region section.

III.9.3.- Levant (F. HIRSCH & M. GAETANI)

The palaeogeography of the Levant is characterised by two W-E oriented embayments. The Palmyra embayment stretches away to the Galilee embayment towards the west, where the Early Ladinian carbonates reach 300 m in thickness (DRUCKMAN *et al.*, 1982). A narrow strip of continental terrigenous sediments extends from Jordan to the Wadi Araba in the Suez Gulf. The Rutbah High separates the Palmyra Embayment from a minor depression to the south in Jordan. Towards the present Mediterranean coast an elongated island emerged as Helez - Gaash high, feeding with clastics a nearby basin. Towards the present Mediterranean, an extensional margin started to develop with a major extensional phase probably during the Carnian (GARFUNKEL & DERIN, 1984; GARFUNKEL, 1998). The faunal association of fucoidal, stromatolitic and lithographic limestones with fish remains, and minor shales with plant remains, shows an high ratio of endemic bivalves, ammonoids, conodonts, crinoids and reptiles, characterising a slightly hypersaline environment, typical for the large shallow Sephardic epicontinental platform Sea (ZAK, 1986). The Tethyan intruders include the ammonoid *Eoprotrachyceras curionii*, allowing correlation to the Tethyan biostratigraphic scale (PARNES, 1986).

The Palmyra embayment was largely filled with dolostones and anhydrites, indicating a wide evaporitic flat facing more open facies to the west, towards the Levant continental margin. The average thickness is of about 300-400 m. almost doubling in the west (BEST *et al.*, 1993). The same facies continues towards the Euphrates Depression through the Bishri Gate, though thickness is less significant. Towards the Aleppo plateau and the Qamlichi Uplift, equivalent of the Mardin Uplift in SE Turkey, the evaporitic facies onlaps Early Palaeozoic rocks (BEST *et al.*, 1993; SAWAF *et al.*, 1993). To the south, the Rutbah High is devoid of Triassic sediments, but we are not certain whether it was active during the Ladinian.

Also to the north in SE Turkey in the Mardin area, drills encountered very variable thickness of the dolostone/evaporitic unit, suggesting significant post-Triassic block-faulting and erosion (TEMPLE & PERRY, 1962; RIGO DE RIGHI & CORTESINI, 1964). The general picture for the northern Arabian promontory is of a passive continental margin, affected by rifting, facing a deep water seaway (YILMAZ, 1993; GARFUNKEL, 1998).

III.9.4.- Arabia and Oman

Most of the North and Central Arabia, as well as part of Iraq and the north of the Gulf, are occupied by a 600 km wide evaporitic platform (Jilh Formation and equivalents; ALSHARHAN & NAIRN, 1997). Towards the Arabian Shield remnants of the more marginal facies with intercalations of fine terrigenous and dry continental flats are preserved (LE NINDRE *et al.*, 1990a). This latter belt may be, however, discontinuous due to later erosion. Towards the Neo-Tethys margin in the Zagros and in the United Arab Emirates gradually a mixed evaporitic /carbonate platform, later dolomitised, eventually developed to reach fully shallow marine conditions in the allochthonous part of the Central Oman Mountains and in the Ras Musandam area, northern Oman. Fragments of the margin, of the slope and oceanic basins are preserved in the nappes of the Central Oman Mountains that have been largely studied during the last 10 years (LE MÉTOUR *et al.*, 1995).

The autochthonous part of interior Oman is instead devoid of Ladinian sediments and a W-E elongated uplift bound the wide Arabian arid equatorial - tropical belt. This uplift might be interpreted as the rift shoulder during the early rifting of the incipient Indian Ocean.

III.9.5.- Somalia and Ethiopia

Fluviatile deposits are referred to the Middle Triassic in the island of Socotra (BOTT *et al.*, 1994).

The evidence of Triassic continental clastic sediments is obtained through boreholes in the Ogaden, as well as by seismic interpretations in the grabens of the Southern Somalia and Ethiopia, both in inland and coastal basins. Because of regional interpretations, these sediments are considered to be Karroo-like in the Mandera - Lugh Basin (BOSELLINI, 1989; ABBATE *et al.*, 1994). In the Ogaden Basin, the Gumburu Sandstone is referred to the Ladinian and to the lower part of the Late Triassic (HENKEL, 1994; HUNEGNAW *et al.*, 1998). It consists of arkoses grading upwards to quartzarenites, deposed in a river channel system (WORKU & ASTIN, 1992). Remains of a volcanic body are supposed to exist in depth near Buq Atoti, in coastal northern Somalia (ABBATE, pers. comm., 1997).

In the Abai River (Blue Nile) graben, continental clastic sediments of Karroo-type are unconformably overlain by the Adigrat Sandstone. Their age is inferred to be Late Palaeozoic or even Triassic. However, their possible Triassic age is highly speculative (BOSELLINI, 1989).

6.- LATE NORIAN (215 - 212 Ma)

M. GAETANI[1]

with contributions of
V. LOZOWSKI, J. SZULC, A. ARCHE, F. CALVET & J. LOPEZ-GOMEZ

I.- MAIN FEATURES

The mapping of the Late Norian (in the substage terminology of the Tethyan Triassic known as Sevatian) was already present in the Tethys project (MARCOUX *et al.*, 1993b). Ammonoid and conodont zones and the bivalve *Monotis salinaria* zone were defined as the base for Late Norian in the Tethys. However, such kind of biostratigraphic tool is hardly of use on the Peri-Tethyan shelves, since although *Monotis* may be scatterly present, like in NW Caucasus (DAGYS, 1963), palynomorphs and ostracods are, at present, the most common biostratigraphic tools used in the marginal areas.

In the Mid European basin, the map was drawn at the level of the upper part of the *Corollina meyeriana* - *Enzonalasporites* zone (palynozone IVC of ORLOWSKA-ZWOLINSKA, 1983) correlatable to the base of Postera-beds in Germany and to the Blue Anchor Formation (=Tea Green Marls) in England. Species of the ostracod genus *Gemmanella* and the species *Lutkevichella keuperea* are also used for regional correlation (CRASQUIN-SOLEAU *et al.*, 1997), even so in many other areas, the attribution to the Late Norian is only tentative. At present, we are in the puzzling situation of having either a very good biostratigraphic scale for the condensed red limestone successions, rich in ammonoid and conodonts, or very good physical (cyclo- and magnetostratigraphy) scales in the lacustrine deposits of the Newark Group and equivalent basins in the Eastern North America and Greenland (OLSEN & KENT, 1996; KENT *et al.*, 1995). Due to the high frequency of magnetic reversal in the Late Triassic, the correlation between the Newark basin and the two Turkish sections of Taurides (GALLET *et al.*, 1992, 1993) cannot be done with confidence. The magnetostratigraphic studies on the Late Triassic rocks are presently in progress in the Mid European basin (G. BACHMANN, pers. comm. 1999), the only Peri-Tethyan area where they have been attempted so far.

As a general consequence, accuracy of correlation may be rather meagre. Often, the map reports facies that are most probably belonging to the upper part of the Late Triassic.

As far as the numerical age is concerned, MARCOUX *et al.* (1993b) adopted an age of 215-212 Ma; the figure of 212 Ma for the top of Norian is confirmed by MENNING (1995). ODIN (1994) quotes an age of 203 ± 3 for the top of the Triassic, but he doesn't give any figure for the top of the Norian. The same data are quoted by the GMW (1998). Accurate geochronological dating of the basalts capping the lacustrine successions of the Newark basin provided an age of 202 ± 1 for the Triassic/Jurassic boundary and the astronomically calibrated scale suggests the figure of 208 Ma for the top of the Norian (OLSEN & KENT, 1996; KENT *et al.*, 1995).

As for most of the Triassic geochronology, in the '90s several proposals have been advanced, yet a coherent and accepted scale is far to be obtained.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

The main kinematic trends outlined during the Late Norian are:

1.- the continuous spreading of the Neo-Tethys, with the anticlockwise rotation of the Cimmerian blocks, led to their eventual collision with the Iranian Spur against Eurasia, starting in the Late Carnian (ALAVI *et al.*, 1997), an age here preferred to the Ladinian suggested by SAIDI *et al.* (1998). The field evidence are thought to be conclusive for the collision of the Iranian Spur during the Late Triassic, and do not support the model with a still open sea-way proposed by KIESSLING *et al.* (1999). This major geodynamic event, responsible of the Eo-Cimmerian orogeny, controlled the evolution of the northern Peri-Tethyan region from the Altaid till North Dobrogea, although far effects of those movements are recognised also in the Mid European basin. In the area here considered, the subduction of the Palaeo-Tethys was completed and the southern margins of Asia and

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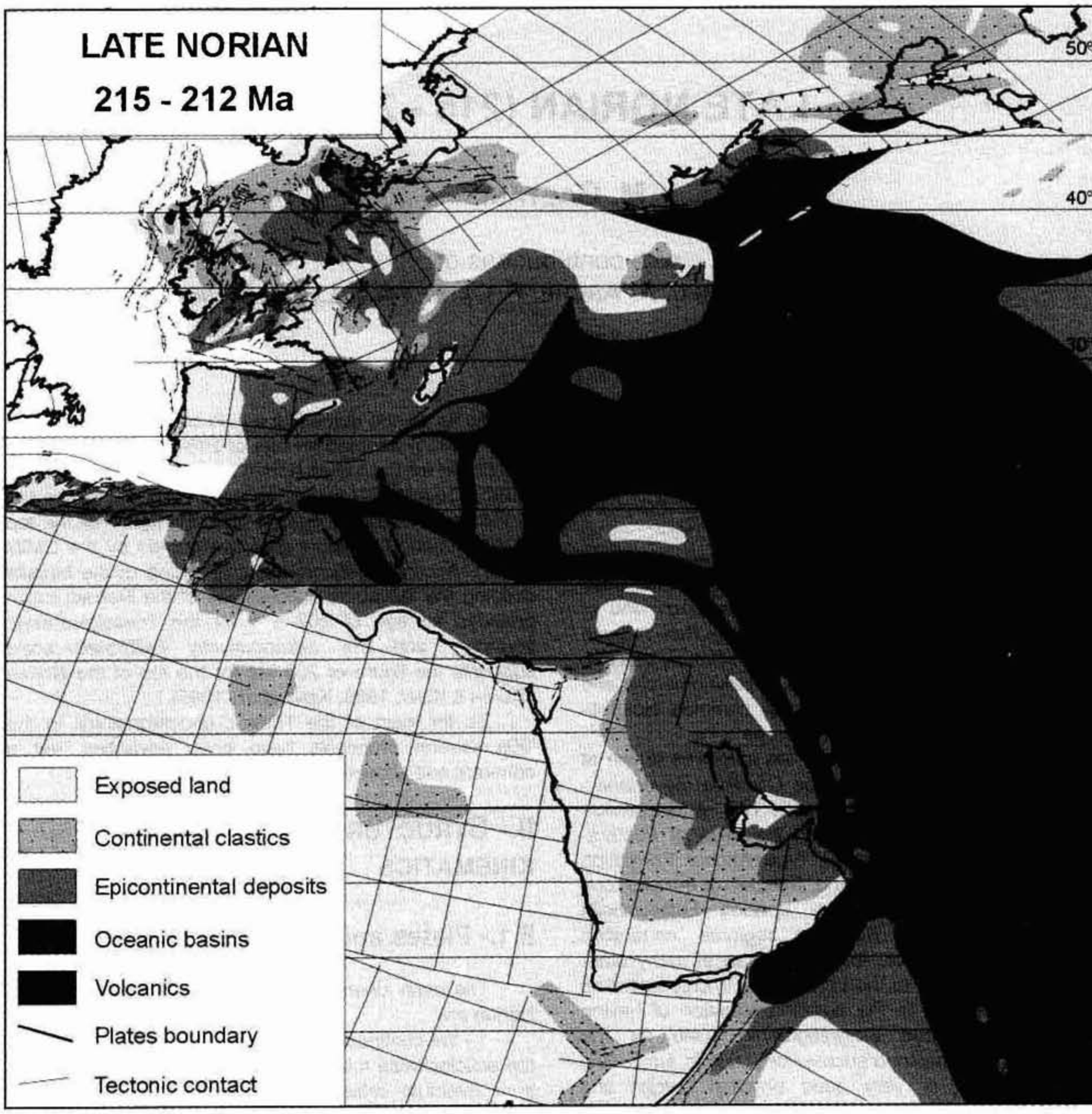

Fig. 6.1: Simplified palaeogeographic map of Peri-Tethyan area during the Late Norian.

Eastern Europe became subsequently the site for the subduction of the Neo-Tethys oceanic crust;

2.- the Pangaea configuration reached definitively its classical Wegener/Bullard configuration during the Carnian (the so called Pangaea A) and then the Central Atlantic started to rift, producing a number of pull apart basins and intense magmatic activity in the Maghreb, especially Morocco, linked to the CAMP (Central Atlantic Magmatic Province) (MARZOLI *et al.*, 1999);

3.- the opening of Neo-Tethys generated also a westwards propagation of the Tethyan branches, further dissecting the margin of Europe. Namely, a deep sea-way was established to the north of the Arabian promontory, resulting in an alignment of basins from the Neo-Tethys to the transcurrent sinistral fault system that allowed the initial rifting of the Central Atlantic and the eastward displacement of Africa.

The sea level tendency seems to be for a generalised regression on the Peri-Tethyan shelves. Most of the marginal basins, sites of evaporitic deposition or having limited connections to the open sea, were controlled by brackish or schizoahaline conditions. The abundance of dolostone in the Norian is also a peculiar feature in the history of Phanerozoic, because such a huge amount of Mg at disposal in the depositional system had no longer such a world-wide counterpart in the rest of Mesozoic and Cainozoic. The dolomitised peritidal platforms, so typical for the Norian, are the very last large early dolomitised carbonate platform in the Earth history.

The main continental blocks accounted for in this time are:

1.- the East European platform in which the Pripyat, Dniepr and Donetz alignment of basins experienced a gentle inversion. To the east of the present 38° E meridian, with the folded Donbass fold belt and the adjacent Karpinsky ridge, the tectonic inversion was active and linked to the Cimmerian collision (STOBVA & STEPHENSON, 1999);

2.- the Precaspian basin;

3.- the system of troughs and basins that border the East European platform and the Baltic Shield from the North Atlantic rift to the North Dobrogea, via the Polish Trough, was much less active and also interested by some tendency to the inversion or at least decreasing of subsidence;

4.- the back-arc basin from North Dobrogea to Caucasus. It received an increasing turbiditic infilling, connected to the active margin to the south;

5.- the Cimmerian orogeny eventually sealed the collage of minor blocks, resulting in the Turan plate. Its southern border became the Kopet Dagh range, on the southern side of which, the Palaeo-Tethyan suture with ophiolite remnants may be observed (STOCKLIN, 1978; SENGOR, 1984; BAUD *et al.*, 1991; ALAVI, 1991; EFTEKHARNEZAD & BERHOOZI, 1991; ALAVI *et al.*, 1997; BESSE *et al.*, 1998);

6.- the microblocks of European affinity, like Moesia, Istanbul, East Pontides, Dzhirula, continued to line the hanging wall of the subduction zone of the Tethys that, from the Norian onwards, formed the Neo-Tethys branch;

7.- the Mid European basin, who should have considerably reduced its connection with the east along the Carpathian Gate because of the ongoing Cimmerian event, whilst its connection with the south widened through the Burgundian Gate. However, such connection was very restricted and the pronounced cyclicity in the sedimentation of the basin appears to be more controlled by regional climatic cycle, rather than by the general eustatic regime. The internal tectonic regime was relatively mild;

8.- the Iberia Block, separated by two tectonic alignments from Western Europe and the northern rim of Gondwana in Africa, followed a history more linked to the Central Mediterranean western shores of the Tethys, than to its Atlantic rim, that entered in its rifting regime;

9.- a major geodynamic event is registered in the Maghreb. Intense transtensional faulting was widespread in the Atlas and Anti-Atlas with Tethyan orientation, oblique to the major transform/transcurrent system that from the Fundy basin in Nova Scotia was developed between Africa and Iberia. On the Atlantic side of the Morocco, a number of asymmetric basins opened, like Argana, Essaouira, Tarfaya, facing the incipient Atlantic, mirroring the equivalent Fundy - Newark, and Richmond type of basins on the North Atlantic coast (MANSPEIZER, 1988);

10.- the Levant and Arabia passive continental margins, with a pronounced asymmetric width of the shelf areas.

II.2.- Palaeoposition of plates and blocks

No available new additional data exists for the APW of major blocks and the kinematic setting of the major plates is that of the Tethys Project (RICOU, 1996). Only the position of Iberia is slightly shifted in comparison to that configuration, according to OLIVET'S model (1996). A research in progress in the Sicani basin, Sicily, has recently obtained an average palaeolatitude of N 18° in the Norian cherty limestones (MUTTONI, pers. comm., 2000).

II.3.- Accuracy

The accuracy of correlation on the biostratigraphic base has been already discussed. As physical stratigraphy is concerned, magnetostratigraphy is at present available only inside Tethys and not on its bordering shelves. The poor dating of many successions hampers correlations based on sequence stratigraphy. Typical is the case of the cyclic sedimentation of the German basin (AIGNER & BACHMANN, 1992; BEUTLER, 1998). Being this basin, for most of the time, not directly connected to the open sea, the oscillations are those of an internal independent basin, depending on regional climate and tectonic pulses and not on the general climatic and tectonic features that controlled the eustatic sea-level curve.

III.- DEFINITION AND DESCRIPTION OF THE DOMAINS

III.1.- East European platform (V. LOZOWSKI)

Of the 3 sub-basins still slightly subsiding during the Ladinian, only the Donetz basin was possibly still slightly subsiding. The presence of terrigenous and coal bearing sediments have been proved in the Bachmuskian depression. To the east of the present 38° E of longitude, Triassic strata are eroded away in the Donbass Fold Belt and the same holds true for the Karpinsky former depression, where flower structures overhanging the neighbouring areas both on the Voronezh and the Ukrainian Shields have been reported.

Likewise, the Pre-Uralian depression was almost filled up during the Late Triassic and only a narrow strip of fine clastic sediments may be tentatively traced between the front of the Ural Range and the emerged East European platform.

The connection between the southernmost part of Pre-Ural depression and the north-east tip of the Peri-Caspian basin is still occupied by terrigenous successions of sandstone and shales.

III.2.- Precaspian basin

Towards the end of the Triassic, the Precaspian basin was towards the end of its 200 Ma long history and evolution. Due to the collision of the Iranian Spur against Asia and the progressive docking of other Cimmerian blocks accreting against Asia, the regional convergence largely stopped or strongly reduced the subsidence in the Precaspian basin. Alluvial plains, swamps, and lakes

covered all the area. Shales and, to a minor extent, fine sandstones and siltstone are the dominating sediments in a prevalently humid climate setting, as suggested by the abundant pollens of hygrophytes and pteridosperms (KUKHTINOV & CRASQUIN-SOLEAU, 1999). Some coal deposits are also reported (ZHIDOVINOV, 1994). General emergence and tendency to the erosion is the constant feature of the end of the Triassic in the area.

III.2.- Border basins at the margin of the EEP

III.2.1.- Scythian platform and Caucasus

The data on the Scythian platform come only from boreholes. They have been recently summarised by NIKISHIN *et al.* (1998a and b). West to east, three main areas may be distinguished: the central-northern Crimea, the Kuban basin in the west Pre-Caucasus and the East Pre-Caucasus basin, subdivided in several sub-basins.

These basins appear in the Late Norian largely linked to the convergence setting that dominate the whole Eurasian margin due to the docking of the Cimmerian blocks. Extensive volcanism has been drilled both in the East Pre-Caucasus where they are very spread, and in the Kuban basin.

In Simferopol area (Crimea), shallow water limestones are tentatively attributed to the Late Norian, and a similar facies is also supposed for the west Pre-Caucasus, though no good biostratigraphic data are available.

The Caucasus continued to be interested by significant tectonic activity. In the NW, the areas of the the tectonic unit of Mount Tchakh, a fairly thick carbonate platform developed through the whole Norian (DAGYS, 1963; SHEVIREV, 1995) up to the Late Norian, overlaid by litharenites still containing *Monotis*. Elsewhere, thick deposits of conglomerates and sandstones are present, mostly fed by the metamorphic basement. They show both evidence of fluvial deposition. The rivers Belaya and Laba, previously sites of emersion and erosion, were eventually covered by a thin veneer of Late Norian sediments (Guseripl) with *Monotis* fauna. In both areas are present evidence of pro-delta and slope features with turbiditic sequences. Due to the sealing produced by late Early Jurassic rocks, it is not possible to ascertain to which extent these conglomerate bodies were deposed during the Late Norian or even later.

III.2.2.- Polish Trough (J. SZULC)

The region between the EEP and the Bohemian massif was strongly affected by the far effects of the Cimmerian convergence, reactivating Variscan lineaments. The homoclinal slope to the north, forming the Lithuanian Embayment, was largely shortened during the Norian. Within the Polish Trough, the mobility is evidenced by erosional gaps between the Norian sediments and older Triassic rocks. A basin-wide unconformity identifies the lower boundary of the Late Norian sequence. This consists of the Zbaszynck Beds, correlative to the "Steinmergelkeuper" of Germany. The map was drawn in the upper part of this unit. The facies are dominated by shallow marine to brackish mudstone flats, with foraminifera and ostracods and, subordinately, by sand flats in which fine-grained siliciclastics were deposed. Caliche horizons are common. The variegated colour of sediments, the braided stream deposits and the caliche horizons points to a semi-dry climate in a subtropical latitude. Notwithstanding the tectonic activity, the topographic relief was smooth, because of the soft nature of sediments that were easily eroded and cannibalised on the relief produced by the movements connected with the Tornquist - Teisseyre fault system. The main connections were with the west, the Fore Sudetic depocentre having been the major subsidence area in Poland.

III.3.- Turan area

Most of the area was emergent or supposed to be, because of the inversion caused by the Cimmerian collision. The Gornyi Mangyshlak was faulted and folded in pre-Toarcian times, according to a NE-SW compression (Fig. 6.2). The dating of the top of the folded succession is not accurate, being generically reported as Carnian - Norian. The only good dating is a Carnian palynological finding (GAETANI *et al.*, 1998), but above this horizon at least 1000 m of sediments follows. At present, we cannot state the age of the last layer involved in the Cimmerian deformation. Very thick alterites, up to 20-30 m thick lies below the coal bearing Toarcian sediments that fill the palaeovalleys. The sealing of the landscape by this new sedimentary cycle occurred only in the Middle Jurassic. All the area was emergent, with, perhaps, the exception of the south Mangyshlak that was still depressed and received some residual sediments. However, dating is poor (MUROMCHEV, 1968). In Tuarkyr, Middle Jurassic continental sandstones lay on the same type of alterites of Gornyi Mangyshlak, even if the younger sediments preserved are Olenekian. The final results support the fact that for the first time it is correct to quote a Turan plate as a coherent slab of continental crust that had a homogeneous behaviour during the following geodynamic evolution. Due to the docking of Central Iran, the subduction zone of Tethys shifted to the south, the Aghdarband volcanic arc ceased its activity, the Palaeo-Tethyan ophiolites were located along the suture between Mashad and Rasht in Iran, and the Kopet Dagh became the actual southern margin of the Turan plate.

III.4.- The North Dobrogea - Küre - Dizi back-arc sea

The continuous subduction of the Tethys, enhanced the development of the back arc basin between the margin of the East European platform and the subduction-related arc. This elongated back arc was extended from North Dobrogea to the Transcaucasus. Their sediments are only poorly preserved as small remnants in Dizi Flysch south of the Caucasus Range, in the Tauric Flysch of Crimea, and in the Istanbul zone, where fragments of Tauric Flysch are possibly present in the Pontides and in North Dobrogea. The latter is the only area where sediments are sufficiently well preserved to allow a more sounded interpretation.

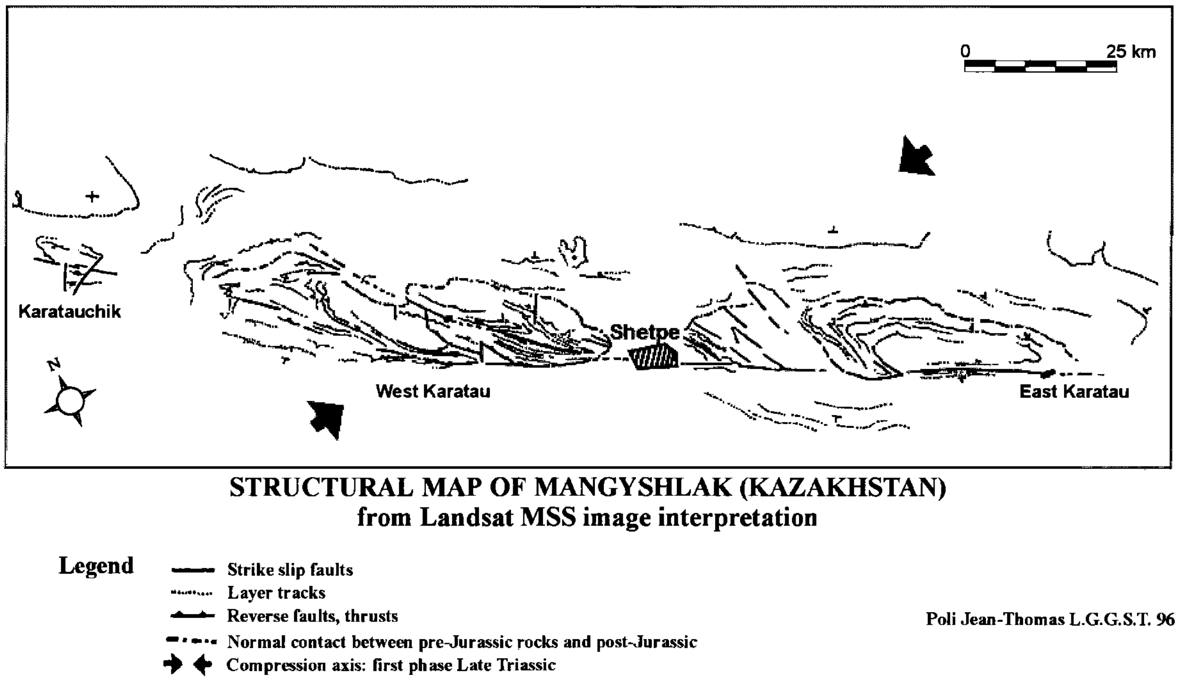

Fig. 6.2: Fold axis and convergence direction of the Eo-Cimmerian deformation in Mangyshlakh, Kazakhstan, Turan plate. The folded Triassic rocks are unconfomably covered by the almost flat-lying Jurassic succession. Courtesy of JEAN THOMAS POLI, UPMC, Dept. de Géotectonique, 1998.

III.4.1.- Dizi Flysch

The upper part of the Dizi Series consists of thin bedded, well sorted, fine grained sandstones and shales, often including tuffaceous material (SOMIN & BELOV 1967; KAZMIN, 1990). Direct evidence of Norian age is however missing.

III.4.2.- Küre basin

The present map adopted the interpretation of USTAOMER & ROBERSTON (1997, fig. 2) and consequently the Küre Ophiolite is considered remnant of the back arc basin to the north of the Palaeo-Tethys subduction zone. The Küre Ophiolite is thought to be covered by Permian - Early Jurassic sediments, but no Norian sediments are known from this unit (USTAOMER & ROBERSTON, 1997).

III.4.3.- Gornyi Crimea

Since long time, exotic blocks embedded in the Tauric Flysch are known from Southern Crimea. The unsettled problem is to establish for sure the age of the base of Tauric Flysch. Moreover, a number of shallow water limestone blocks embedded in the flysch and containing Late Triassic (Norian - Rhaetian) has been known for a while (MOISSEEV, 1932). More recently DAGYS (1963, 1974) and SHEVIREV (1990) described the brachiopod and ammonoid fauna originated from the blocks, some of them including also *Monotis salinaria*. The origin of these blocks is to be searched not far from the present outcrops, possibly linked to a huge collapse along the active thrusting that interested the northern margin of the back-arc basin.

III.4.4.- North Dobrogea

Stratigraphy and facies pattern have been recently summarised by GRADINARU (1995) and SEGHEDI (2000).

The North Dobrogea shows, in a now telescoped nappe structure, a transect from shallow water carbonates to deep-water turbidites. To the east, the carbonate ramps is still existing, but the new fact is the onsetting of turbiditic sedimentation to the south west suggesting that also in the North Dobrogea, the Cimmerian convergence was in progress and the inversion with thrusting of the northern margin of the basin was going on. In particular, the siliciclastic turbidites are fed from western sources, in the Tulcea zone and in the Macin zone where evidence of erosion and non-deposition was obtained in boreholes (SEGHEDI, 2000).

In the map, a connection is supposed to exist still with the Polish Trough, Another possibility could suggest the onsetting of the Eo-Cimmerian deformation temporary closing the link.

III.5.- Microblocks of European affinity

1.- Dzhirula and Khrami in the Transcaucasus, slabs strongly deformed during the Hercynian Orogeny with evidence of intrusive and sedimentary rocks of Palaeozoic age, probably emergent during the Norian (ADAMIA & LODKIPANIDZE, 1989).

2.- The Eastern Pontides contains a metabasite - phyllite - marble unit, tentatively considered of Permo-Triassic age (OKAY & SAHINTÜRK, 1997).

3.- In the Western Pontides, where the Istanbul Zone or Unit is preserved (OKAY, 1989; OKAY & SAHINTÜRK,

1997), sediments assigned to the Norian are mixed siliciclastic and carbonatic breccias. However, the presence of rocks of Late Norian age is doubtful.

4.- Moesia. The Moesia block and its proxy Fore-Balkan, previously largely submerged by a shallow water sea during the Late Norian, were also interested by the onsetting of the Cimmerian convergence. Evidence of local emersions with carbonatic breccias (Chelyustnitsa Formation) are quarried as ornamental stone in the Forebalkan, interbedded with shales and marls of brackish environments. However, it cannot be demonstrated that this deposits persisted up to the Late Norian or that, instead, most of the area was already emergent and subject to the progressive deformation that led to the Cimmerian unconformity, which is of importance in the regional petroleum system (TARI *et al.*,1997; GEORGIEV, 1996)

III.6.- Middle European basin

The Polish extension has been already described. The bulk of the basin is in Germany, where it is represented by the Postera Beds, that are considered in the Germanic literature as Lower Rhät (BEUTLER, 1998). Basically, two depositional domains may be distinguished. The proximal one, with more clastic influx, especially towards the Baltic Shield, from where more abundant supply was originating, forming a wide deltaic system (the Postera Sandstone), suggesting a more humid climate to the north. This wide fan divided the basin into an eastern part, more restricted, and a western part largely developed under the North Sea. Conversely, to the south the clastic influx was less significant, with a wide bay corresponding to the onset of the Paris basin, in which the rivers from Massif Central and Armorica formed a wide delta gradually passing to a schizohaline lagoon with shales and dolostones. Getting more to the south, more arid climate prevailed and more frequent evaporites may be found. On the contrary, in the centre of the basin a pelitic facies dominates and the dominant brackish water conditions are characterised by ostracod and charophytes faunal and floral remains. Soft bottom bivalves like *Unionites posterus* are also fairly spread. Thin layers with marine foraminifers attest to temporary links to the south through the Burgundian Gate and to the east through the East Carpathian Gate.

III.7.- British and Ireland Isles

The emergent London - Brabant, Welsh and Grampian positive structures closed towards the west the Mid European basin. The very low relief of these lands allowed establishing wide flats, fine grained, silty to clayey, grey-green in colour (Blue Anchor Formation). To the south, this unit is thicker, lithologically heterogeneous in its upper part, thus recording the transition to the shallow marine environments that marked the beginning of the Rhaetian transgression (WARRINGTON & IVIMEY-COOK, 1992).

Continental clastic sedimentation characterised most of the northern areas in Scotland and Ireland, although west of Scotland biostratigraphic evidence for the Norian is lacking. To the south-west, instead (Western Approaches and Celtic Sea), transitional gradually passing to marine muddy facies are recorded. Biota were fairly rich (miospores, *Euestheria*, vertebrate remains and footprints). The very gradual lateral transition of the facies suggests that tectonic activity was fairly reduced during this time.

III.8.- Iberia (A. ARCHE & M. GAETANI)

On the stable Iberia, the facies distribution is very simple. A wide carbonate ramp gradually overlaps from the east the eastern margin of the Iberian massif under a tectonic regime of mature thermal subsidence. The best development is in the Catalanid ranges where it reaches the 30-70 m in thickness for the Late Norian. In the map, such facies has been interpreted as indicative of open marine, even though the generalised dolomitisation and the local presence of some evaporitic intercalation suggests the tendency towards hypersaline environments (SOPENA *et al.*, 1988; ARNAL *et al.*, 1999).

To the north in the Pyrenees and to the south in the Alpujarrides of the Betic Cordillera, shallow water limestones are widely spread, testifying to the presence of two more open marine corridors through these areas. Thus, the tectonic regime for the Late Norian continued to be of thermal subsidence on the Iberian microplate, bounded to the north and to the south by active rifting domains. To the south, an important strike-slip component was also present.

On the western side of Iberia, the ongoing rifting that lead to the Atlantic opening in the Jurassic is active. A set of basins with asymmetric grabens and half grabens faced west (RASMUSSEN *et al.*, 1998; PALAIN *et al.*, 1977). These basins, mostly filled by mudstones locally inclusding halite pseudomorphs, might be partially considered of Late Norian age. Similarly, on the southern coast in Algarve, a rifted basin filled with fine alluvial sediments (Grès de Silves; PALAIN, 1977) opened towards the transtentional depression linked to the sinistral movement between Iberia and Maghreb.

III.9.- Gondwanan margin

III.9.1.- Maghreb

The palaeogeographic map of the Late Norian is fairly homogeneous at the scale of the NW Africa, displaying clear trending from the mostly evaporitic to marine environments in Tunisia, to continental environments in the south of the Saharan platform in Algeria and in depocentres in Morocco (COUREL *et al.*, 2000). This is, however, only a snapshot. If we broad the picture over a slightly larger span of time i.e. to the whole Norian, it appears that the eastern evaporitic environments moved westwards over time, in a clearly established retrograding trend in the Norian in the northern part of the Saharan platform of Algeria. In the same area, isopachs of the evaporitic formation (S4 in the local terminology) of Late Norian/?Rhaetian age, indicate the existence of a fairly homogeneous salt basin, extending E-W, cutting the preferentially NE-SW Carnian age depocentres. E-W fracturing in Tunisia is though to have controlled basalt flows (Chott region), which are evidence of a rifting phase

on the South Tunisian margin. A E-W trend marked the Norian deposits in Tunisia and Algeria which may be viewed as a vast rather homogeneous platform, opening westwards to the marine domain, without significant structuring and with evaporitic to schizoaline dominating environments.

The situation was markedly different in Morocco. The east-west trend towards more distal environments in the east, still prevailed over much of the eastern Morocco. On the opposite, in the Atlantic coastal domain, to the south of the High Atlas, dominant current directions were towards west, like in the Argana basin. In the coastal domain along the Atlantic, normal faults prevailed along N-S and NNE-SSW fracturing directions, both onshore and offshore. The faults bounded grabens filled with clastics derived from the east (High Atlas and Meseta) and with thick evaporites in the Doukkala, Essaouira and Tarfaya basins. Blocks were sometimes tilted eastwards. Volcanic activity is also widespread, associated to the rifting activity, linked to the earlier extension movements along the future Central Atlantic.

III.9.2.- Libya and Egypt

The region may subdivided in two parts, separated by an alignment of positive structures, like the Gargaf High that remained under erosional setting. To the south, two major internal basins, the Murzuk and Al Kufrah basins, contain the so-called Post-Tassilian Nubian Sandstones, which may span from the Late Permian to earliest Cretaceous. In the map, the presence of Norian sediments is only tentative, suggesting persistence in the deposition.

To the north, the more articulated situation of the Middle Triassic of Tripolitania was over and the area simply represents the southern margin of the wide evaporitic platform and continental flat already described for Tunisia. The western Desert area of Egypt was mostly under erosional conditions. In the Matruh basin and Gebel Rissu basin, increasing links to marine shelf developed gradually during the Late Triassic. The restricted carbonates, evaporites and sands of the Fadda Formation are attributed to this age (KEELEY, 1994). The Delta area was never penetrated below the Cretaceous. Consequently, the inference that marine sedimentation occurred in that area during the Triassic is purely conjectural. The Sinai and Wadi Araba areas also largely emerged. If some sedimentation occurred, it was lately eroded during the Cretaceous (GARFUNKEL, 1998).

III.9.3.- Levant

The Late Norian is not well documented in the area, because of the general regressive trend and because of the important Early Cretaceous erosion that largely removed older sediments from the marginal areas. The onshore palaeogeography of the Levant is still characterised by a SW-NE oriented embayment. The Palmyra embayment was largely filled with dolostones and anhydrites, indicating a wide evaporitic flat facing more open facies to the west, towards the Levant continental margin. The tectonic activity allowing the subsidence of the embayment was associated with normal faulting. The later inversion with thrusting along the southern front of the Palmyrides might have occurred along early listric faults (LOVELOCK, 1984). The Late Norian was palaeontologically documented only in the coastal plain of Israel in the Ga'ash borehole, where about 200 m of dolomitised oo- and pelmicrites were penetrated (Shefayim Formation). Towards the Euphrates Depression, on the Bishri gate, evaporitic flat overlapped sandstones (Mulussa F unit), are interpreted as derived from erosion on the Rutbah High, some of them of possible Late Norian age (DE RUITER *et al.*, 1995). Towards the Aleppo Plateau and the Qamlichi Uplift, equivalent of the Mardin Uplift in SE Turkey, the evaporitic facies onlaps Early Palaeozoic rocks (BEST *et al.*, 1993; SAWAF *et al.*, 1993). To the south, the Rutbah High is devoid of Triassic sediments.

Also to the north in SE Turkey in the Mardin area, drills encountered very variable thickness of dolostones and slope breccias, suggesting significant post-Triassic block-faulting and erosion (TEMPLE & PERRY, 1962; RIGO DE RIGHI & CORTESINI, 1964). The general picture for the northern Arabian promontory is that of a passive continental margin affected by rifting, facing a deep water sea-way (YILMAZ, 1993; GARFUNKEL, 1998). Instead, towards the present eastern Mediterranean, in the Late Triassic, a passive margin largely oriented parallel to the present coast was progressively structuring. Faultings like the Helez fault delimiting the Ga'ash high, has a total throw of 1 km for the whole Triassic. The onsetting of a rifted basin of some tens of km in width, allowed the widening of the distance between the African continental margin and the Erathostenes High (GARFUNKEL, 1998).

III.9.4.- Arabia and Oman

The wide platform of the Arabia facing the Tethys formed a depression deeply entrenched into this side of the African continent. Towards the south-east, on the contrary, the emerged Qatar Arch separated the coastal Oman, where the present nappe system of the allochtonous telescopes the facies pattern, not allowing a detailed palaeogeographic reconstruction.

In Saudi Arabia, a very wide delta apron of silicilastics flowing towards NNE is present, in which both braided-river and alluvial meandering plain have been detected (LE NINDRE *et al.*, 1990a, fig. 20). In the outcrop belt, where for the first time the Minjur Formation was described (POWERS *et al.*, 1966; SHARIEF, 1986), it is possible to follow the gradual passage from coarser to finer sandstones in a decreasing energy environment, leading to intercalations with marshy, brackish flats, that are revealed in the subcrop of Central Saudi Arabia. Further to the large, mostly towards Kuwait, Gulf and Central Iraq basin, marine coastal flats are developed, interfingering basinward with the shallow carbonates. The Zagros High continued to be emergent, acting as shoulder of the renewed rifting of the Neo-Tethyan margin. In the outcrop sections of the United Arab Emirates, the Minjur Formation (Ghalailah Formation *in* ALSHARHAN & NAIRN, 1997) shows alternance of ferruginous quartzarenites with grey marls (ALSHARHAN, 1993). The top is eroded in some areas, with Upper Norian rocks that may be presently missing. In the nearby mountains of Oman, the Ras Musandam and, especially, the Djebel Akhdar expose calcareous walls formed in a shallow water peritidal platform, hundreds m-high, so giving an

alpine view to the landscape (PILLEVUIT, 1993). An intermediate continental plain is supposed to have existed between the shallow carbonate platform and the emergent interior. However, no positive evidence is preserved. The allochtonous of Oman preserve a complex pattern of Triassic facies from the base of the talus to the pelagic, which have been intensively studied in the last years (for a synthesis see PILLEVUIT, 1993).

In inner Oman and Yemen, all the area was emergent for the rifting that was actually propagating between the Indian plate and Africa, eventually led to the opening of the Indian ocean in the Jurassic (LE MÉTOUR *et al.*, 1995).

III.9.5.- Somalia and Ethiopia

The tri-radial system of grabens: Abai River, Ogaden and Mandera - Lugh, continued its evolution, passing during the Late Triassic from the pre-rift stage with the Karro-like sediments to the syn-rift sediments, represented at its base by the Adigrat Sandstone (BOSWORTH, 1994, HANKEL, 1994). In the Ogaden basin, the best known because of petroleum exploration, the Gumburo Sandstone is gradually replaced by the medium to coarse quartzarenites of the Adigrat Sandstone, testifying to the transition from braided to meandering river system (WORKU & ASTIN, 1992). The age is poorly constrained through palynology, Late Triassic and earliest Jurassic.

The presence of Late Triassic continental clastic sediments is also supposed through seismic interpretations and few boreholes in grabens of the southern Somalia, both in inland basin both in coastal basins (BOSELLINI, 1989; ABBATE *et al.*, 1994). Remains of extrusive volcanics are reported near Hafun, Somalia, at the base of the Adigrat Sandstone (BOSELLINI, 1989). In the Abai River (Blue Nile) the Adigrat Sandstone crops out, but its age is also here poorly constrained, perhaps already Jurassic since its base (BOSELLINI, 1989).

7.- LATE SINEMURIAN (193 - 191 Ma)

J. THIERRY[1]

I.- MAIN FEATURES

The Late Sinemurian map was not present in the Tethys Programme; but, a schematic and simplified sketch showing the main palaeoenvironments and palaeogeographies of the Tethys has been constructed (FOURCADE *et al.*, 1996).

In the Sinemurian ammonites zonal scheme for the West European and Mediterranean Jurassic (CORNA *et al.*, 1997), the selected Obtusum Zone (Late Sinemurian; Lotharingian), falls at the base of the substage. Subsequently, the "Lotharingian" (*sensu gallico*) corresponds to the "Late Sinemurian" (*sensu anglico*).

Correlations are very good all over NW Europe because ammonites became more and more ubiquitous since the Hettangian. On and after the Middle Lotharingian begins a faunal crisis which will amplify till the Pliensbachian (CARIOU *et al.*, 1985). Some specific taxa characterise the Tethys area and its north and south borders, but precise correlations are possible, so far to the east on Moesia, Crimea and Caucasus.

The time interval represented on the Late Sinemurian map is as far as possible restricted to the Obtusum Zone; but, sometimes it includes the overlaying Oxynotum Zone, or the underlying Turneri Zone, or both, because of similar facies. Where ammonites are missing, especially in carbonate platforms of South Europe, central west and south Tethys, brachiopods (ALMÉRAS *et al.*, 1997) and larger benthic foraminiferas (BASSOULLET, 1997a) are used, but the biostratigraphic resolution is wider. Therefore, the data recorded on the map may span on all Late Sinemurian, sometimes on all Sinemurian or Early Liassic. Appearing during the Late Triassic in the Tethyan realm but still rare, calcareous nannofossils associations are the dating bioevents used at a substage precision for the deep carbonate - hemipelagic oozes facies of the Early Jurassic of the North Europe and Mediterranean areas (DE KAENEL *et al.*, 1996; GARDIN, 1997). In addition with rare dinoflagellate cysts (RIDING & IOANNIDES, 1996; FAUCONNIER, 1997), ostracods (BODERGAT, 1997) and smaller benthic foraminiferas (RUGET & NICOLLIN, 1997), they allow fairly good datings and correlations especially for shales and clay facies, in outcrops and subsurface data of the Sub-Boreal realm. Therefore, the facies and palaeoenvironments recorded on the map may correspond to the Late Sinemurian or the whole Sinemurian.

In continental, coastal plain or shallow marine environments (West European platform, Polish and Kharkov basins, Precaspian and Iranian areas, intracratonic Saharan - African - Arabian - Nubian basins and Atlantic margin platforms), spores and pollen are the only possibility for tentative correlations, as often as at the stage precision, or sometimes for the whole Early Liassic.

The palinspastic reconstruction is interpolated according to the Norian and Toarcian Tethys maps (RICOU, 1996).

The Sinemurian has not yet yielded a verified magnetostratigraphy, but the majority of the early Late Sinemurian, including the Obtusum Zone, appears to be dominated by reverse polarity (STEINER & OGG, 1988; YANG *et al.*, 1996) which falls between 197-199 Ma (GRADSTEIN *et al.*, 1995).

As far as the numerical age is concerned, the lower and upper boundaries of the Sinemurian stage are respectively 200-204 Ma and near 191 Ma (ODIN, 1994). Subsequently, because of the general agreement for an equal duration for each ammonite zone, the time slice concerned by the Late Sinemurian map may falls between 194-197 Ma with a bar error of more or less 4 Ma.

Considering the above data, the Late Sinemurian map illustrates the palaeogeography near 193-191 Ma. Such age is in harmony with other time scales (GRADSTEIN *et al.*, 1994, 1995) where the lower boundary of the Late Sinemurian is near 198 ± 4 Ma and the upper boundary near 195 ± 4 Ma. It is consistent with the age of 194.1 ± 0.6 Ma given for the boundary between the Early Pliensbachian (Early to Middle Carixian) Ibex and Jamesoni Zones in Canada (THOMSON & SMITH, 1992).

II.- STRUCTURAL SETTING AND KINEMATICS

II.-1. Plates and blocks accounted for

Among the five major plates which composed the Pangaea, only two of them, mostly continental, are concerned:

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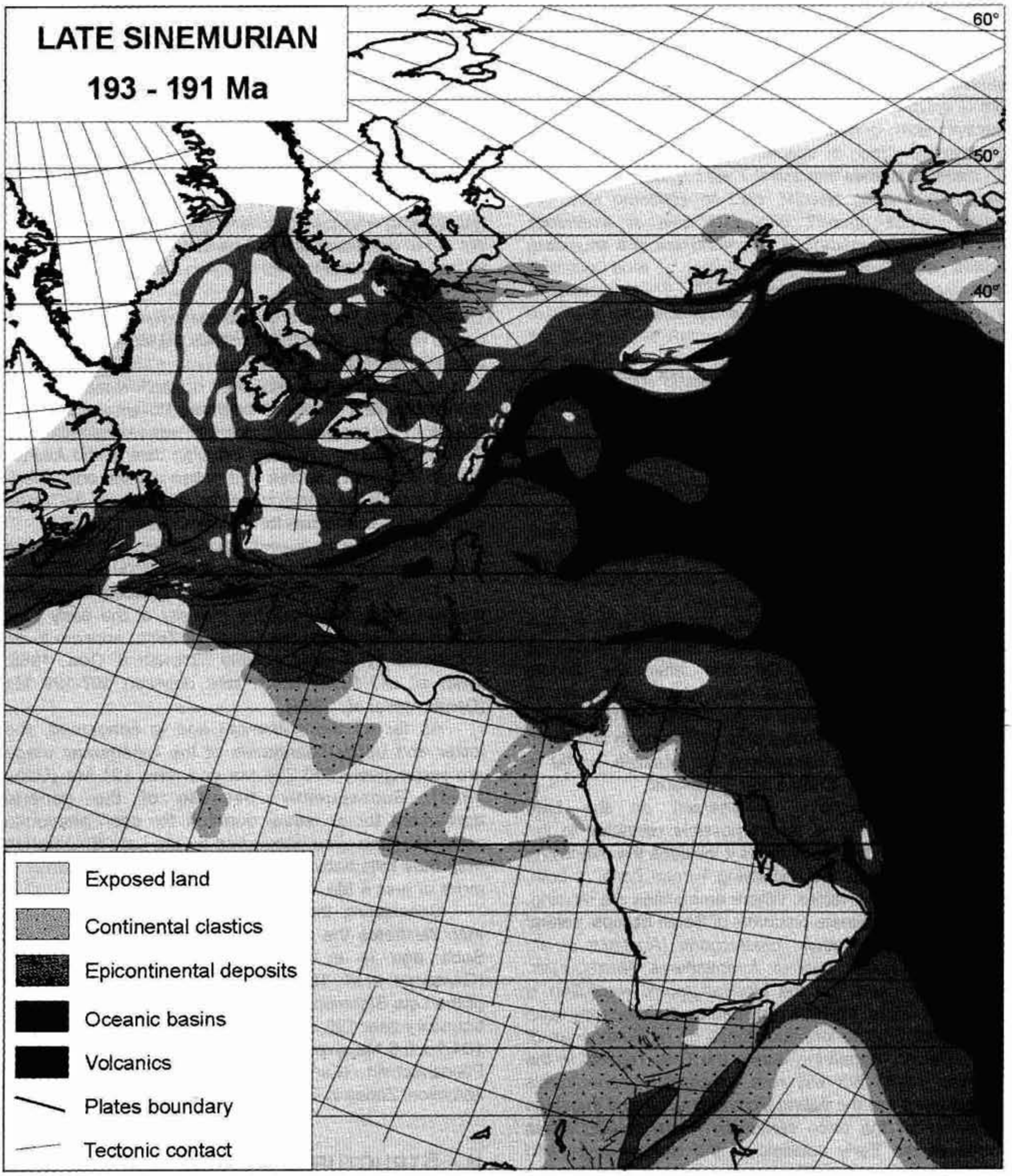

Fig. 7.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Late Sinemurian.

Laurasia for the north Tethyan areas, with the Iranian spur henceforth coupled south of the Caspian - Turan plate since the Norian (Ricou, 1996); west Gondwana for the south Tethyan areas, with the west border of the east Gondwana (Indian plate) still linked, facing the Arabia - Somalia block. Considering a rotation axis on the northern foreland of the present day Scandinavian peninsula during all the Jurassic, the general tendency is a gentle anticlockwise rotation of Laurasia which drives southwards its Tethyan margin. At the same time, Gondwana does the same and its SE parts (Arabian platform and Indian plate) shift southwards. In compensation, the West European borders and the still linked Greenland and NE American parts shift slightly northwards; the North African areas do the same, in spite of the beginning of the Central Atlantic rifting.

Four main kinematics trends outline the Early Liassic times:

1.- since the Late Norian - Rhaetian, far from the east in Trans-Caspian areas onto North Dobrogea to the

west, the geodynamic evolution along the east border of the north Tethyan regions is controlled by the "Early Cimmerian" event. Westward effects of these movements are recognisable with difficulty in the Mid and West European basins. As in Late Triassic, the south margin of east Europe is still the site of subduction of oceanic crust;

2.- the Pangaea, which has reached its configuration since the Norian - Rhaetian, begins to break-up. The Central Atlantic starts to rift (STEINER *et al.*, 1998; ZIEGLER *et al.*, 2000), producing several pull apart basins along the present day west Morocco coast. Active extension plays between West European and Saharan cratons, all over Maghreb and its boundary with the Iberian block: the "Maghreb Jurassic transform zone" (RICOU, 1996). Active extension also occurs in Ligurian surroundings while passive thermal subsidence controls the evolution of major basins in several regions of the North European craton (North Sea, Paris and Polish basins);

3.- the opening of the Tethys continues to generate a westward propagation of Tethyan branches. The future Vardar and Bükk oceanic furrows with several connected deep sea areas (Budva and Pindos - Olonos troughs) dissect the margin of Europe; they are the interconnected results of the E-W alignment of Neo-Tethys Triassic basins along several transcurrent sinistral faults, the rifting of the Central Atlantic and the eastward displacement of Africa. Thus, West Tethys appears as a complex patchwork of blocks, more or less clearly isolated: Moesia, Ticza, Austro-Alpine, north Alpine and south Alpine, Pelagonian, Karst and Puglia. Similarly, the long and deep Pamphylian basin separates the north of the Arabian promontory from the Gavrovo - Taurus block till the Ionian basin;

4.- the border of the Iran - Arabia - India craton is a large continental passive margin widely opened to the east, facing the Tethys ocean.

The main blocks accounted for in the Late Sinemurian time can be listed:

1.- the East European platform is part of the Laurasia; the active tectonic inversion linked to the "Early-Cimmerian" collision has quite completely obliterated the Pripyat - Dniepr - Donetz basins alignment, except the small continental Kharkov basin, the only remnant part of the Ukrainian areas with passive thermal subsidence;

2.- the Precaspian areas and the Turan plate similarly undergo a tectonic inversion as a Cimmerian consequence; they are in continuity with the emerged Stavropol High with isolated continental deposits;

3.- the narrow back-arc basin, from North Dobrogea to Caucasus, through south Crimea, receives an abundant detritic flysch-like infilling, related both to its active south margin and the huge emerged areas at the north;

4 - the Iranian spur, henceforth included in the north Tethyan domain, represents the south border of the Great Caucasus trough with the Trans-Caucasus blocks;

5.- the microblocks of European affinity, like the complex Moesia and East Pontides, are underlined by the subduction zone of the Tethys;

6.- a system of troughs and basins lines the east European platform and the Fenno-Scandian shield, from North Dobrogea to the Ergesund - Danish basin through the East Carpathian Gate and the Polish Trough. The latter, with continental infilling, much less active than during Triassic because of a decreasing subsidence, separates the Bohemian massif from the north shields;

7.- as a consequence of constant marine invasions on lowlands, associated with a regional feeble tectonic regime, the Mid European areas widen their connections to the north (Franconian and Hanover basins), to the south (Schwabian platform, Valais trough and Dauphinois basin) and to the west (Paris basin). The sedimentation is controlled by regional constraints and general eustatic regime, even though some deposits show an evident cyclicity which may be of climatic influences;

8.- the North European domain appears as a large triangle-shaped archipelago, inserted between the Fenno-Scandian shield on the east, and the Greenland shield on the west; its base is widely opened to the south (North Sea, narrow troughs and basins on the British Isles), while its head is a very narrow channel (Viking Graben) which allows communications with the Boreal realm ("Greenland - Scandinavia seaway");

9.- the Iberian Block is still off the Western Europe; it is shifted along the tectonic alignments of the future Bay of Biscay rift and separated from North Africa by the complex north margin of Gondwana (part of the future "Hispanic seaway"). Its Atlantic rim (Lusitanian basin), enters in its rifting regime, opening communications between the Boreal realm until the Maghreb areas and the originating Atlantic domain;

10.- the Atlasic domain of the whole Maghreb registers an intense transtensional faulting which widely spreads in Tethyan orientation, from Morocco to Tunisia, oblique to the major transform/transcurrent Atlantic system. On the Atlantic side of the Moroccan Meseta, a number of asymmetric basins open (Essaouira - Agadir, Tarfaya), facing the incipient Atlantic;

11.- the High and Lower Saharan platform (Oued Mya, Dahar, Chotts and Tataouine) registers an active N-S extension in intracontinental and marginal basins assumed to be connected with the Murzuk basin;

12.- the present day shoreline of Libya and Egypt, from the Pelagian areas to Levantine façade, coincides with a long passive margin with supposed normal faulting. Endoreic basins (Kufra, Erdis, Lakhla) or basins in connection with the marine domain (Dakhla) extend on the Libyan - Egyptian part of the African craton;

13.- the Levant and Arabia - Iran areas are passive continental margins with pronounced asymmetric width of the shelf parts. Showing eastward embayments, the shelf is very narrow and probably controlled by N-S normal faults along the Levantine front; it is larger on Arabia and Iran. Both would be connected by mean of an E-W platform below fluctuating shallow marine conditions controlled by a normal faulting subsidence which extends from Syria - Lebanon, between the Mardin high, the Hamad uplift and the Rutbah high;

14.- the "African corner" (Somalia, Ethiopia) is occupied by a large and complex basin mainly structured by two families of normal faults. The first one is parallel to the present day east Somalian coast, coinciding with the future opening between Africa and India. The second one is orthogonal with a NW to SE orientation. The result is a complex of continental grabens (Blue Nile, Berbera - Borama, Ogaden and Anza basins) which surround

narrow marine areas in southward connection with the Malagasy province, but still isolated from the Tethys.

II.2.- Palaeoposition of plates and blocks

The kinematics setting of the plates is extrapolated from the Tethys maps (RICOU, 1996) because no available palaeomagnetic data exist. The position of Africa and North America was obtained directly from the fit along the east Coast anomaly (KLITGORD & SCHOUTEN, 1986). There is no computed new position of blocks and the palaeolatitude grid refers to the palaeomagnetic pole of BESSE & COURTILLOT (1991).

The plate model evolution, with a basin still opened north of the Turan plate, proposed according to palaeomagnetic studies carried out on the Early Permian to Late Triassic formations of the Scythian and Turan plates (LEMAIRE, 1997; LEMAIRE *et al.*, 1997; LEMAIRE *et al.*, 1998a and b) is in disagreement with the classical reconstruction (RICOU, 1996); these blocks are attached to the southern border of Eurasia until the end of the Triassic.

The position of the Iberian plate is extrapolated from the kinematics model of OLIVET (1996), deduced both from the Atlantic Ocean and the Bay of Biscay openings, and the evolution of the Pyrenees (VERGES & GARCIA-SANZ, 2000). Iberia is slightly west shifted and rotated in comparison to the Tethys maps. The Corsica - Sardinia block is NE shifted facing the Mediterranean border of France, while the Balearic area remains linked to the Iberian block, according to their facies and palaeo-environments similarities, respectively with Alpine and Iberian - Betic areas (FOURCADE *et al.*, 1977). In comparisons with the Tethys Programme maps and geodynamics of the Gulf of Lion (VIALLY & TRÉMOLIÈRES, 1996), the consequence for all the Jurassic Peri-Tethys maps is an enlarged marine area between the Alpine - Provence - Corsica - Sardinia on the east, and the Betic - Iberia - Balearic Islands on the west.

The Moesian microplate position is a major problem on all the Jurassic maps. In spite of evident facies and palaeoenvironments similarities which suggest Tethyan connections with the Great Caucasus and Crimea, an option to move it to the east is not satisfying; the result is a strange remaining wide space to the west, indenting the European margin and strongly contrasting with the Pontides area. However, there is tectonic reasons to move Moesia to the east. In a general transpressive regime, the eastern position of Moesia allows to have a dextral movement for its displacement. If instead it is moved to the west (BANKS & ROBINSON, 1997), the movement became sinistral, contrary to the general pattern between Laurasia and Gondwana. In order to avoid a non-proved motion of Moesia during the whole Jurassic as consequences of the opening and closing of the Magura, Vardar and Dobrogea - Crimea furrows, its position during Sinemurian has been arbitrarily fixed a little more to the east than on Tethys Programme maps. Such a reconstruction remains in agreement with the palaeolatitudes data of Moesia and the Eurasian palaeolatitude curve (SURMONT *et al.*, 1991; TCHOUMATCHENCO *et al.*, 1992); it does not correspond to the reconstructions which open a wide "Meliata ocean" south of Moesia and Austro-Alpine - Carpathians micro-plates (STAMPFLI, 1993, 1996; STAMPFLI *et al.*, 1998a, b and c, 2000). Whatever it be, the Peceneaga - Kamena fault, which separates North and South Dobrogea in Romania, is a major tectonic feature between Moesia and the north Tethyan margin; it is interpreted on the Jurassic maps as related to an hypothetical fault which underlines south Crimea, connected with the transform faults of Caucasus.

II.3.- Accuracy

The magnetic sequence with numerous inversions proposed for the Hettangian and Sinemurian of the Paris Basin (YANG *et al.*, 1996) and Austria (STEINER & OGG, 1988) is not yet really verified and no synthetic and reliable scale exists, even though some data are calibrated with an ammonite zonal scheme. Therefore, the accuracy of the palinspastic reconstruction is linked to the classical Atlantic fits.

The accuracy of correlation based on the biostratigraphic data has been already discussed: it is generally reliable and with a high or fairly good resolution.

Correlations based on sequence stratigraphy are emphasised by the very good biostratigraphic dating of many successions, overall in NW Europe (RIOULT *et al.*, 1991; GRACIANSKY *et al.*, 1993; GRACIANSKY *et al.*, 1998a, b; STEPHEN & DAVIES, 1998; HESSELBO & JENKINS, 1998; DUMONT, 1998; VAN BUCHEM & KNOX, 1998) which plays as the type area to construct a "Jurassic Sequence Chronostratigraphy / Biochronostratigraphy Chart" (HARDENBOL *et al.*, 1998). During quite all the Jurassic, in the several independent analysed basins, the marine oscillations registered depend both on the regional climate and tectonic pulses, and on the general climatic and tectonic features that control the eustatic sea-level curve.

II.4.- General comments

The break-up of the Pangaea has still begun during the latest Triassic - Earliest Jurassic, but rifting is not at the same step all over the Peri-Tethys areas. Still in an early phase in the future Atlantic Ocean and its dependencies (Atlasic domain), or on North and East European platforms (North Sea, Polish Trough), it is in high activity in the Ligurian domain which has not yet oceanic crust. Compressive tectonics along the eastern Moesia - Scythian - Crimea - Great Caucasus - Alborz region has just ended in Hettangian ("Early Cimmerian" orogenesis), as witness of the collision between Eurasia and several terranes (western Eastern Pontides, Trans-Caucasus, Alborz; OKAY *et al.*, 1994; ROBINSON *et al.*, 1996).

As a whole, the north Peri-Tethyan borders of the emerged areas are underlined by coastal-shallow marine terrigenous facies, while on the south, till the latitude of France, shallow marine carbonate deposits dominate on platforms and ramps. The collage of the Iranian block to the Turan plate has enlarged the continental conditions

and the clastic facies all over the NE Peri-Tethyan border.

The Late Sinemurian map takes place during the major Liassic transgressive episode ("Ligurian Cycle", *sensu* JACQUIN & GRACIANSKY; 1998) which began in Late Triassic (Norian/Rhaetian boundary) up to the tectonically enhanced "Early Cimmerian unconformity" (Aalenian/-Bajocian boundary), and which peak transgression is registered near the Early - Middle Toarcian boundary. The map illustrates one of the significant Jurassic marine invasions (second order transgressive - regressive facies cycles) on the Laurasia and Gondwana cratonic borders. These two massive Pangaean blocks, still in majority below continental conditions, surround the V-shaped Tethys ocean, closed to the west ("Mediterranean *Seuil*"; VRIELYNCK *et al.*, 1996) and largely opened to the east ("Central Tethys"). During the transgression, most of the marginal platforms and basins which had previously limited connections to the open sea till the latest Triassic - earliest Liassic (Rhaetian - Hettangian) and which filled with evaporitic or brackish deposits, are progressively drowned. The marine transgression is much more important on the north Peri-Tethyan areas than to the south where evaporitic facies are still well developed and marine shelves still narrow.

Marine connections are proved when a series of transgressions led to faunal unification between Boreal domain and the platforms of NW Europe through the "Greenland - Scandinavia seaway" (DORÉ, 1991, 1992); none exists through the totally emerged Russian platform. No marine connections are documented between European - Maghrebian areas and Western Caribbean Tethys through the incipient Atlantic Ocean and the "Hispanic seaway".

III.- DEFINITION OF DOMAINS

III.1.- Russian platform: Volga - Ural - Donetz - Ukraine

The Russian platform is a huge emerged area which undergoes strong erosion since the Late Triassic - Early Jurassic "Early Cimmerian" deformations. The successive Late Carnian - pre-Norian, Rhaetian - Hettangian boundaries and Hettangian orogenic phases may be related to possible collision with Trans-Caucasus, Iran, Western and Eastern Pontides terranes (NIKISHIN *et al.*, 1998a and b). The Russian platform is in continuation to the west with the Fenno-Scandian shield and to the east with the Ural highs; it extends to the south until the Ukrainian - Stavropol shields and the Precaspian high (VINOGRADOV, 1968). The Hettangian - Sinemurian are not documented, except in the Early Liassic continental deposits of the small Kharkov - Donetz basin related to a possible regional and reduced passive thermal subsidence, without conspicuous faulting (STOVBA *et al.*, 1996; VAN WEES *et al.*, 1996; STEPHENSON *et al.*, 2000). The present day limits of the Jurassic sequence are certainly erosional; the original depositional area of the Donetz area would be wider (ULMISHEK *et al.*, 1994).

III.2.- Turan plate

The Turan domain corresponds to the vast present day area which extends east of the Caspian Sea to the Aral Sea, between the Kopet Dagh depression to the south, and the Russian platform - Precaspian basin to the north. It includes the central Karakum, the north Ustyurt depression, the Mangyshlak swell - peninsula, the south Mangyshlak depression, the Tuarkyr and Kara Bogaz highs, the Aral basin and the south margin of the Precaspian basin. The Turan plate is definitively cratonised by the "Early Cimmerian orogeny" (Iranian - Turan plates convergence) during the Norian - Rhaetian, causing a gentle folding and tilting of the Triassic successions (LYBERIS *et al.*, 1998; LYBERIS & MANBY, 1999). A weak rifting regime would be still present during Early Jurassic onwards, contemporaneous with the opening of the Great Caucasus trough which is attributed either to a back-arc extension (NIKISHIN *et al.*, 1998b) or to the propagation of a rift (RICOU, 1996). However that may be, the Jurassic development of the Turan domain occurred in a NE-SW extensional framework, driven by a normal fault trending and a tilted block system (LYBERIS *et al.*, 1998; THOMAS *et al.* 1999) which progressively install from south (Kopet Dagh range and depression) to north (central Karakum platform - depression).

Generally, the Jurassic deposits unconformably overlie the Triassic. Intracontinental - limnic to fluvial deposits (conglomerates, sands, silts and clays, sometimes with bauxites-like alterites or coal bearing; Daghestan), supposed to be Sinemurian (or Toarcian?), fill the depressions and bottom of the palaeovalleys of an irregular palaeorelief. North Ustyurt, South Mangyshlak, Kara Bogaz, Central Karakum and other denudation areas situated further to the east provide the clastic input. Alternative models are proposed for the direction of the vast fluvial systems which feed the lowlands dissected into isolated ridges (PANOV *et al.*, 1996; VOLOZH *et al.*, 1997). On the one hand, the Turan - Middle Caspian fluvial network is assumed to have a sub-latitudinal orientation; on the other hand, sub-longitudinal orientation is demonstrated on the north part of the Caspian region and the SE part of Turkmenistan through Uzbekistan to the north Aral region.

III.3.- Scythian platform - Crimea - Black Sea - Caucasus and Precaspian areas

During the whole Jurassic onwards, the evolution of the Scythian - Black Sea - Pontides - Great Caucasus domain is governed by a succession of extensional and compressional phases, controlled by the tectonic activities along a north-dipping subduction zone (NIKISHIN *et al.*, 2000). The continental Iranian terranes accretion, accompanied by a succession of back-arc compressions associated with a dextral transpressive regime ("Early Cimmerian" orogeny), caused the inversion of the Triassic rifts. The Scythian platform, Precaspian, south Caspian and Alborz terrane areas are a vast emerged region which undergoes erosion, as a south prolongation of the Russian platform and Stavropol - Ukrainian shields.

In North Caucasus, the Triassic sedimentary cover was removed; the Hettangian - Sinemurian are generally not documented, possibly due to a thermal uplift of the Black Sea - Great Caucasus north shoulder (NIKISHIN *et al.*, 1998a and b; ERSHOV *et al.*, submitted). The Great Caucasus is first a continental rift-like sedimentary basin in a back-arc environment (Rhaetian - Hettangian); therefore, it opens during the Late Sinemurian - Early Pliensbachian, as proved by ammonite faunas and foraminiferas in shallow marine terrigenous deposits in west Caucasus. A deep-water trough is underlined by slopes (Kuban and Terek basins) which fill with fine-medium detrital "flysch-like" sediments yielding ammonites (PANOV & GUSHIN, 1987). It prolongs eastward in the south Caspian and Kopet Dagh basins (BRUNET *et al.*, submitted) where the extensional phases of the "Early Cimmerian" orogeny led to the deposition of the continental - fluviatile coal-rich clastic "Shemshak formation"; the south Caspian basin is east closed because the Iranian plate deposits are fed from the north (DAVOUDZADEH & SCHMIDT, 1983; OTTO, 1997).

Affected also by the Rhaetian - Hettangian - Early Sinemurian orogenic phase, south Crimea is first an uplifted region (type area of the "Cimmerian" events), then a subsiding area drowned by the transgressions, until the Late Sinemurian - Pliensbachian onwards. Shallow-water clastics with scarce carbonates developed on the margins of the south Crimea trough while its slopes and bottom fill everywhere with clastic clayey to sandy deposits, sometimes turbiditic (PANOV *et al.*, 1994); such facies and environments extend until the present day Gulf of Odessa, Dobrogea, and NE Black Sea in a back-arc basin which undergoes extensional movements, possibly as soon as Sinemurian (ROBINSON *et al.*, 1996; ROBINSON & KERUSOV, 1997).

During the Early Jurassic, continental terrigenous series of the north slope of the Great Caucasus and the Precaspian basin represent a complex of a large fluvial system which extends onto the Scythian - south Russian - Turan plates. A compromise between the two interpretations (PANOV *et al.*, 1996; VOLOZH *et al.*, 1997) of the orientation of the fluvial network is illustrated on the Late Sinemurian map.

Data on the Dzhirula massif in the Trans-Caucasus and uplifted areas in Pontides back-arc are very scarce. These emerged areas, south of the Great Caucasus trough and Black Sea belt have been reduced compared to their extension on the Tethys maps because it is difficult to admit that very large crustal blocks must almost completely disappear as it is to day. Narrow marine gates separate them from the Alborz terrane headland and the complex Moesian domain.

III.4.- Teisseyre / Tornquist zone - Moesia

The Teisseyre / Tornquist zone forms the easternmost Permian - Mesozoic NW European area where several basins install during the Jurassic, superimposed on the crustal boundary between the west and east European cratons. The relationships between the sedimentary infill of these basin, their tectonic framework and the Palaeozoic history of the basement are clearly demonstrated (DADLEZ, 1997; NARKIEWICZ, 1997). Following a NW-SE trending orientation, the north Danish - Ergesund basin, the Mid Polish basin/trough, the so-called "east Carpathian gate" (today partly blinded by the Carpathian range) and Dobrogea are concerned.

At the onset of the Jurassic (Hettangian - Sinemurian), these basins are north bordered by the uplifted units of the Fenno-Scandian - Ukrainian shields and Scythian platform which are the land areas supplying clastics to the basins. The sedimentary characteristics indicate a strong climatic turn-over if compared to the Triassic; a subtropical semi-dry Late Triassic climate changes into a cooler and more humid Early Jurassic one (termination of the red beds deposition). The shallow marine sandy-clayey north Danish - Ergesund basin is connected with the monotonous north Germany - Hanover epicontinental open sea shaly deposits. The Mid Polish basin is continental, south bordered by the Bohemian massif, a large coherent and permanent land mass during the whole Jurassic (MALKOVSKY, 1987; ZIMMER & WESSELY, 1996), from which clastics are shed into the adjacent areas of Poland and Germany.

During Early Liassic (Hettangian - Sinemurian), the geometry, facies and palaeoenvironments deposits of the Mid Polish basin fill, indicate tectonic and sedimentary similarities with features characteristic of rifting basins: a uniform subsidence pattern is guided by the continuous activity of syn-sedimentary bounding faults of a transverse asymmetric-half graben basin; intrabasinal uplifts (Wielkopolska High) are active during the Hettangian - Pliensbachian interval (DADLEZ *et al.*, 1995; HAKENBERG & SWIDROWSKA, 1997; LAMARCHE *et al.*, 1998; LAMARCHE, 1999). At the beginning of the basin opening, the sedimentation is clastic dominated in varied freshwater deltaic - alluvial plain to lacustrine and swampy - lagoonal environments (PIENKOWSKY, 1991; DADLEZ *et al.*, 1998; MAREK & GRIGELIS, 1998). Short incursions of the sea, underlined by brackish to shallow marine sandy-shaly facies are limited to small areas in NW Poland; dated as Late Hettangian, Sinemurian and Pliensbachian they allow to recognise transgressive - regressive cycles which can be correlated with those of Western Europe (FELDMAN-OLSZEWSKA, 1997a).

The complex Moesian micro-plate is composed of two lithospheric blocks, the Moesian platform, facing to the north the Dobrogea furrow, and the Thracian/ Rhodopes massif, facing to the south the Tethys ocean (TCHOUMATCHENCO *et al.*, 1989; SAPUNOV & TCHOUMATCHENCO, 1990). During the Early Jurassic, the Moesian platform slowly subside as marine conditions extend, whereas Thracian/Rhodopes massif is constantly uplifted. These blocks are E-W separated by a multitude of horsts and grabens, mostly today included into the Balkanides Range. Moesia is separated from the east European platform by Dobrogea, a marine flysch-like deposition trough, disposed as the west prolongation of the Caucasus - Black Sea - Crimea furrow; it is supposed to be tectonically connected with the east Carpathian gate and the Polish Trough. Alike in several areas of the north Tethyan margin, the "Early Cimmerian" compressive events are registered. Above the Early Cimmerian unconformity, the sedimentary facies and thickness rela-

tions of the Early Jurassic deposits clearly show the development of a south-facing passive margin.

Hettangian - Toarcian coarse terrigenous deep-water facies are restricted to North Dobrogea; Sinemurian is dated by ammonites, spores and pollen (GRADINARU, 1993; SEGHEDI, 2000). The Bulgarian side of the Moesian platform begun to subside slowly since the beginning of the Hettangian, accompanied by a progressive disintegration into E-W trending grabens and horsts, bounded by normal faults which dissect the previously uplifted areas (SAPUNOV & TCHOUMATCHENKO, 1990; TCHOUMATCHENCO & SAPUNOV, 1994; TARI *et al.*, 1997). In continuation of more or less developed Late Triassic - earliest Liassic continental clastic facies, gradual marine ingressions enter several grabens (Sevlievo, Mihajlovgrad, Izdremets, Veliko - Tarnovo, Provadija) and furrows (Nish-Trojan trough). The horsts (Vidin, Vraca, Pleven and east Balkans) remained emerged lands which provide clastics for the marine deposits dated by ammonites, brachiopods, smaller foraminifers and spores and pollen (SAPUNOV & TCHOUMATCHENCO, 1987, 1990; SAPUNOV *et al.*, 1988, 1991; TCHOUMATCHENCO & CERNJAVSKA, 1989).

III.5.- Western Europe platform

The West European platform is a vast epicontinental sea, dotted with a lot of more or less extended and variable emerged blocks. The "Ligurian transgressive-regressive cycle", very well and everywhere biostratigraphically constrained, can be taken as a canvas to describe the Early Jurassic palaeogeographic evolution of Western Europe (GRACIANSKY *et al.*, 1998a). The sea level variations are mainly related to the several phases of two tectonic events that affected the southern parts of the European cratonic areas but interfere too on the vast nearby northward areas. The first one, the Ligurian rifting, on the future alpine range, leads to the opening of the Ligurian Tethys (LEMOINE & GRACIANSKY, 1988; DUMONT, 1998) and brings an influence upon its surroundings (Paris basin, Schwabian platform and Provence - Corsica - Sardinia complex). The second one is related to the opening of North Atlantic, responsible of major events: the onset of the rifting of the Lusitanian margin (SOARES *et al.*, 1993; CANÉROT *et al.*, 1995); the intracontinental rifting that separates Iberia and Europe on the future Pyrenean range, which interferes on the evolution of the Aquitanian basin (VERGES & GARCIA-SENZ, 2000); the setting of the Iberian and Betic "rift-like" systems (CANÉROT, 1989, 1991; SALAS *et al.*, 2000; VERA, 2000).

Within the Liassic transgressive episodes, from the Early Hettangian to the Late Sinemurian upwards, the sea-level rise leads to a progressive drowning of quite all the remnant highs, and the overstepping of the basin margins which enlarged the epicontinental sea; the history of the transgression can be divided into high resolution and very well dated depositional sequences which are correlated quite all over the Western Europe (GRACIANSKY *et al.*, 1993, 1998a and b; HESSELBO & JENKYNS, 1998; STEPHEN & DAVIES, 1998). The areas of shallow-water carbonate of the Hettangian and Sinemurian, which extended northward, toward the London - Paris basin and adjacent areas, are progressively dominated by silts and shales sedimentation, sometimes with interbedded kerogenous deposits (FLEET *et al.*, 1987; BESSEREAU & GUILLOCHEAU, 1994; HANZO & ESPITALIÉ, 1994; MORTON, 1993). Subsequently, the carbonate platforms, still well developed at the onset of the Late Sinemurian (Aquitaine basin, Provence platform, Iberian and Betic ranges) begin to shift southward near the end of the Late Sinemurian and during the Pliensbachian, until the Toarcian.

In the future Alpine areas (Ardèche, Dauphinois basin, Valais trough, Briançonnais), a tilted blocks framework delimits ridges and basins where the subsidence is enhanced by the extensional effects of the rifting activity. In the Paris basin, the marine transgressions progress towards its west and SW borders, however without direct communication with the Aquitaine basin and without reaching the Cotentin peninsula which remained only opened onto the English basin; shallow sandy-limy deposits underline the emerged areas (Armorican and London - Brabant massifs). The Early Liassic tectonic and sedimentary evolution of the flexural intracratonic Paris basin is overall the result of a long term thermal subsidence as a consequence of the Permian extension (GUILLOCHEAU *et al.*, 1999; ROBIN, 1997; ROBIN *et al.*, 2000); but, several synsedimentary tectonic trends indicate the superimposition of a short-term tectonic component due to intraplate deformation related to the several steps of the Ligurian opening (ROBIN *et al.*, 1996, 1998).

Large connections install with the Schwabian and Franconian platforms which undergo a marly to shaly sedimentation; shallow sandy-limy deposits underline the west border of the emerged Bohemian massif, onto the Polish Trough - Ergesund - Danish basin. Although there is few evidence for synsedimentary tectonics, the axis of several basins coincide with the trace of the Permian fault systems (MALKOWSKY, 1987; SCHRÖDER, 1987; ZIMMER & WESSELY, 1996).

Guided by a weak rift activity, the transgressions progress northward and invade numerous inland and offshore basins, all over the United Kingdom (Celtic Sea, Western Approaches, Channel, Hebrides, Minches, Irish Sea, Wessex, Moray Firth; ANDREWS *et al.*, 1990; CALLOMON & COPE, 1995; EVANS, 1990; HAMBLIN *et al.*, 1992; HESSELBO & JENKYNS, 1995; JACKSON *et al.*, 1995; JENKYNS & SENIOR, 1991; MORTON, 1987, 1989, 1990, 1992a and b, 1993; MORTON & HUDSON, 1995; RAWSON & WRIGHT, 1995; STEPHEN *et al.*, 1993; TAPPIN *et al.*, 1994), North Sea (Broad Forteen, Sole Pit, Central Graben; CAMERON *et al.*, 1992; GATLIFF *et al.*, 1994; PARTINGTON *et al.*, 1993a and b), West Netherlands (VAN ADRICHEM-BOOGAERT & KOUWE, 1997) and North Germany (BRAND & HOFFMANN, 1963; WEITSCHAT & HOFFMANN, 1984; KÖLBEL, 1968; BOIGK, 1981; KOCKEL, 1995). Dating and correlations are supported both by biostratigraphy (ammonites for inland outcrops and dinoflagellate cysts - spores and pollen for offshore boreholes) and sequence stratigraphy. These basins fill with marls and shales, somewhere organic rich; sandy input frequently underlines the emerged areas. Local and regional synsedimentary faulting is everywhere evident. Numerous unconformities and gaps suggest that the basin extension would be more important; the Early Liassic sequence may have been

deposited and removed later, prior or during the Middle Jurassic, as a consequence of the North Sea thermal doming (UNDERHILL & PARTINGTON, 1993)

The drowning of the areas between Shetland platform - Faeroe high - Greenland and Norway (Viking Graben) opens the way between boreal and subboreal - Mediterranean realms through the North Sea and across Great Britain, the Paris basin and the Franconian - Schwabian platforms. Sequence stratigraphy investigations (STEEL, 1993; THOMAS & COWARD, 1996; LEPERCQ & GAULIER, 1996) on the Jurassic successions of the northern North Sea suggest that the sequentiality is not merely caused by eustatic variations; it probably includes a major contribution from subsidence-rate variations, driven by the regional post rift thermal subsidence in relation with the previous Triassic rifting event.

III.6.- Maghreb (Morocco - Algeria - Tunisia) - Saharan areas

In the Sinemurian, North Africa shows a strongly differentiated palaeogeography inherited from the latest Triassic - Early Liassic aborted rifting of the Atlasic domain (ELMI, 1996; PIQUÉ *et al.*, 1998; KAMOUN *et al.*, 1999): large emerged lands (Saharan craton and Moroccan Meseta); isolated shoals and/or emerged uplifts (intra-Tellian shoals, Medenine High); inner platforms changing through time into outer platforms, carbonated ramps and more or less isolated lagoons (Middle and High Atlas; Ksour Mountains and Saharan Atlas; Tunisian Dorsale, North-South Axis and Chotts basin); large sabkhas (Essaouira - Agadir - El Jadida and Tarfaya basins; Oran High plains, Tlemcen domain, Lower Sahara - Oued Mya - Dahar - Ghadamès basins; Tataouine basin) and continental basins (Oued Mya and Ghadames basins; Murzuq basin). During the Hettangian - earliest Sinemurian, a shallow marine transgression gradually flooded over North Africa, driving to the "initiate carbonate platform" episode (ELMI, 1996; SOUHEL *et al.*, 2000). These initial extensional basins which are opening until the Early Mesozoic, are part of the Atlantic - Tethys transfer zone; they are progressively invaded by the Early Liassic transgression. Good datings and correlations are given by ammonites, brachiopods, foraminiferas, dinoflagellate cysts, spores and pollen (PEYBERNÈS, 1992; EL HARIRI *et al.*, 1996; ELMI *et al.*, 1998; POISSON *et al.*, 1998; LACHKAR, 2000; SOUSSI *et al.*, 2000) which allow to well constrain the sequence stratigraphy framework and the basins infilling events chronology (SADKI, 1992; SOUHEL *et al.*, 1998, 2000).

Near the end of the Early Sinemurian, the Northern African margin, especially the "Atlas" areas, undergoes a strong subsidence (ELLOUZ *et al.*, submitted) and deepening, in a rifting context (VIALLY *et al.*, 1994), depending on an oblique W-NW - E-SE transtensive trending. The tectonic activity (LAVILLE *et al.*, 1995; El KOCHRI & CHOROWICZ, 1996; ZIZI, 1996; PIQUÉ *et al.*, 2000; CHOTIN *et al.*, 2000), influenced by the evolution of both Atlantic and Tethys oceans, is directly related to crustal thinning (BRACENE *et al.*, submitted). This dynamic regime, which produces tilted blocks, half grabens and syn-sedimentary faults, was initiated in Hettangian and Early Sinemurian and continues during the Late Sinemurian and the Early Carixian (POISSON *et al.*, 1998). In the Great Kabylia, the rifting which occurs in a progressive manner from the south towards the north, affects at the beginning the external domain of the chain, therefore the median part (CATTANEO *et al.*, 1999). In the Rif basin and foreland, sedimentary facies dated by ammonites, foraminiferas and brachiopods, indicate a progression of rapid subsidence and extensional tectonic on a passive margin, related to a crustal stretching as a consequence of Tethyan and Atlantic rifting episodes (FAVRE *et al.*, 1991; EL HATIMI, 1991; MEHDI *et al.*, 1994; BOUTAKIOUT & ELMI, 1996). All over the Maghreb, the initial platform is progressively broken into numerous basins (ELMI, 1996) with relatively deep hemipelagic to pelagic areas (High Atlas Trough, Saharan Atlas, Tunisian Trough), slopes and platforms (Middle Atlas and Oran High plains, south border of the Saharan Atlas and Chotts areas), shoals and islands (Medenine High). A tilted-block model can explain the persistence and synchronism of these environments; the shallow carbonated deposits of the crest of tilted-blocks can show coral build-ups correlated with deeper marly sediments.

At the end of the Sinemurian ("Lotharingian event") until the Early Carixian, a major sea level rise, coeval with the tectonic activity, leads to the establishment of an "initial platform carbonate" regime (ELMI, 1996) on the previously emerged or sabkha areas (Oran High Plains, Atlas troughs margins, Ouarsenis intra-Tellian shoals). A strong dynamic and environmental decoupling is evident between the Atlas domain and the future alpine northern sectors. The beginning of the "mosaic" episode (ELMI, 1996) is often marked by the development of silicified or nodular limestones and *ammonitico rosso* facies (PEYBERNÈS *et al.*, 1996; ELMI *et al.*, 1998; SOUSSI *et al.*, 1998, 2000) along the margins (Ksour Mountains, Djurdjura, Calcareous Ridge of the Rif) and in the centre of the basins (Saharan Atlas, Tunisian Trough). Coeval with the peak transgression, the "partitioning" episode (ELMI, 1996) reaches the northern areas (Tlemcen and Tell domains) in the Early Domerian; even at least, the High Plains were locally invaded by the open sea.

A narrow fringe of shallow marine environments separates the High Atlas trough from the emerged Saharan platform. It enlarges eastward and extends in the wide areas of the paralic to evaporitic and continental deposits (the so-called "continental intercalaire") of the Lower Sahara, Oued Mya, Dahar, Ghadames and Tataouine basins. The large evaporitic deposits suggest an arid climate during the Late Triassic - Liassic - Early Dogger times (LEFRANC & GUIRAUD, 1990; BUSSON & CORNÉE, 1991). Platform carbonates are developed in the Tunisian dorsale and North-South Axis, until the Chotts basin (PEYBERNÈS *et al.*, 1990; SOUSSI *et al.*, 1991a and b; KAMOUN *et al.*, 1999; SOUSSI *et al.*, 2000). The latter is structured into horsts and grabens which undergo active submeridian extensions along E-W normal faults related to the Africa - Eurasia divergence accompanying the Jurassic rifting (BOUAZIZ *et al.*, 1998). A weak but continuous and homogeneous subsidence spread over most of Tunisia and the Pelagian platform, until the Hammamet, Gulf of Gabes (HLAIEM *et al.*, 1997; PATRIAT *et al.*, submitted) and offshore Malta basins (BISHOP & DEBONO, 1996); in the latter, the Tethyan distension is marked by

neptunian dykes which develop along a fault scarp (BOUILLIN *et al.*, 1999). As far as possible the evaporitic supratidal to intertidal deposits of the Tunisian - Libyan Saharan platform (Jeffara and Tatatouine basins) are refereed to the Late Sinemurian, their sedimentation is likewise controlled by a N-S extension (BARRIER *et al.*, 1993; BOUAZIZ *et al.*, 1996a and b, 1999). In west Libya (east Ghadames basin; BELHAJ, 1996) marginal marine to non marine deposits are refereed to Hettangian - Sinemurian, while in SE Libya, the Murzuq basin, northward opened on the Ghadames basin and Tunisian areas, fills too with continental Nubian sequences which Jurassic age is assumed.

On the Atlantic side of Maghreb, the Essaouira and Tarfaya basins are encroached on the emerged Moroccan Meseta promontory and Saharan craton; they result from a NW-SE trending extension, initiated from the south and migrating to the north, which induces the formation of westward-dipping half-grabens (MEDINA, 1994, 1995; LABASSI *et al.*, 2000; CHOTIN *et al.*, 2000). Up to the evaporitic facies of the Late Triassic - Early Liassic "salt province" series, syn-rift deposits grades from coastal plain to tidal-flat and inner platforms. Fairly good biostratigraphic data (ammonites, brachiopods, foraminiferas, calcareous algae and coccoliths) in outcrops and offshore boreholes allow a precise dating (PEYBERNÈS *et al.*, 1987; DU DRESNAY, 1988; DE KAENEL & BERGEN, 1993; BROUGHTON & TRÉPANIER, 1993; MEDINA, 1994; MORABET *et al.*, 1998). Differential subsidence and the proximity of the emerged areas influenced the facies distribution; a northern part, dominated by shallow platform carbonates and marls, with coral build-ups or dolomitic facies, differentiates from a south part with rather abundant terrigenous input (conglomerates, sandy-silty limestones). No connection exits between the western toe of the Atlas rift and the Essaouira basin, despite the facies and environments are very similar on both sides of the emerged Moroccan Meseta peninsula.

Related to the first abortive rifting episode known all around the future Central Atlantic and Atlasic domains, some of the basalts intrusions are possibly Rhaetian - Early Hettangian (Lower Sahara) and ante-Rhaetian (Saharan Atlas). Falling near 200-208 Ma (WILSON, 1997; GUIRAUD, 1998; WILSON & GUIRAUD, 1998), this plume related event precedes the Central Atlantic opening. It is not reported on the Sinemurian map which represents a nascent narrow Atlantic sea, yet with shallow to deeper marine environments, but without oceanic crust.

III.7.- Egypt - Sudan - Libya

The Early Jurassic, from Sinai Peninsula to the Gulf of Sirt, is often poorly documented: outcrops are rare and the series are mainly known from inland or offshore boreholes; the sediments, overall deposited in continental to coastal plain or shallow shelf environments, provide few reliable biostratigraphic data.

The classic palaeogeography is a more or less wide passive margin, gently northward dipping, in a parallel alignment with the present day Mediterranean coast line; the several facies make a belt shape area ("unstable shelf"; KERDANY & CHERIF, 1990), encroached along the emerged African craton ("stable shelf"). This one shows rather complex structures, affected by normal faulting and the shear system generated by the plate movements in the Tethyan realm; from Rhaetian to end-Bajocian, the Gebel Rissu and north Gulf of Suez-Sinai basins are the only individualised syn-depositional elements of the "unstable shelf" (KEELEY & WALLIS, 1991). They are occupied by coastal plain and shallow shelf of low relief environments which grade northward to the deeper marine Tethyan sea. Continental sediments belong to several basins separated by uplifted emerged areas in relation with fractures. The depositional framework will not change greatly through the entire Jurassic.

Because precise biostratigraphic documents are frequently missing, no more precise dating than Early Jurassic stage or substage is often possible. Moreover, the tectonic features are too controversial overall on the unstable shelf despite the latter is considered as a rift complex: on the one hand, E-W normal faults, parallel to the present day coast line, control the sedimentation from the North Sinai to the Cyrenaica platform (MOUSTAFA & KHALIL, 1990; GUIRAUD & BELLION, 1996; MOUSTAFA *et al.*, 1998; GUIRAUD, 1998; GUIRAUD & BOSWORTH, 1999); on the other hand, NE-SW fractures, orientated in a parallel direction to the Trans-African lineament, gradually developed as regional fault-bounded basins, producing intra-basin segmented faults blocks which formed separate half grabens (KEELEY & WALLIS, 1991; ANKETELL, 1996; EL-HAWAT, 1996; SMITH & KARKI, 1996; KEELEY & MASSOUD, 1998; AYYAD *et al.*, 1998). Hence, depending to the chosen model, the deduced palaeogeography changes following the alternative solutions (THUSU *et al.*, 1988; SCHANDELMEIER *et al.*, 1997; KEELEY & MASSOUD, 1998; GUIRAUD *et al.*, 2000).

The Liassic, involving the Sinemurian, is either supposed (JENKINS, 1990) or totally missing (KEELEY *et al.*, 1990; KEELEY & WALLIS, 1991; KERDANY & CHERIF, 1990) in north Sinai (Gebel Maghara). Often referred to the so-called "Nubian Sandstones", the facies yields coarse clastics with minor red silts and shales, of fluvial to coastal plain environments; they progressively northward grade (extreme north and offshore Sinai) to low energy shallow shelf and inner to outer platform carbonates and shales (JENKINS, 1990; KEELEY *et al.*, 1990; KEELEY & WALLIS, 1991; KEELEY, 1994). The abundance of continental clastics is probably the result of a low sea level coupled with tectonic uplifts, dissected by faults during an Early Jurassic rifting (AYYAD *et al.*, 1998). A similar facies succession extends westward in the Nile Delta and the extreme North-Western Desert, and the Cyrenaica coast; showing mostly continental to coastal plain environments, it is controversially considered as Liassic (Sinemurian to Pliensbachian - Toarcian) or younger (ABDEL AAL *et al.*, 1990; HANTAR, 1990; KERDANY & CHERIF, 1990), on the basis of palynologic data. Hence, the North-Western Desert is either in continuation with the continental Dakhla basin or an emerged area, alike the North-Eastern Desert .

The Sirt basin is reputed to be devoited to Jurassic deposits (MASSA & DELORT, 1984). Therefore, the west and east stable parts of Libya are separated until at least late Middle or early Late Jurassic by a large uplifted

emerged area which extends from the Tibesti and Gargaf high to the Cyrenaican arch.

On the stable areas in Egypt, Sudan and Libya, the Dakhla, Kharga, Al Kufra, Erdis and Lakla basins fill with continental and fluviatile clastic sediments; probably controlled by tectonics, the sedimentation is supposed to initiate during Early Liassic or, at least, Late Liassic - Dogger, or Late Jurassic (KLITZSCH & WYCSISK, 1987; HERMINA, 1990; KLITZSCH, 1990; KLITZSCH & SQUYRES, 1990; KLITZSCH & SCHANDELMEIER, 1990).

III.8.- Levant (Israel - Lebanon - Syria - Jordan)

The palaeogeographic model is a Levantine passive margin which had opened by rifting at the beginning of Jurassic (BEST *et al.*, 1993; SAWAF *et al.*, 2000; WALLEY, 2000). Probably underlined by normal faults, the narrow slope of the continent and the sea floor spreading zone would be not very far west to the present day coast-line. Several data support this interpretation: from east to west, the facies belts show continental to littoral facies and a very thick shallow carbonate sequence which is greater than can be accounted for by the Jurassic eustatic variations even allowing isostatic adjustments; the rare but unquestionable deeper facies and palaeoenvironments lie immediately west to the shallow facies; the clastics input in Levant is ascertained to come mainly from the wearing down of the emerged Arabian platform in the south and SE; volcanic intrusions witness aborted intracratonic rifts at the beginning and the end of the Jurassic (MOUTY *et al.*, 1992; LAWS & WILSON, 1997). Therefore, the Levantine carbonated Jurassic sequence probably illustrates a deposition which keeps pace with a margin undergoing post-rifting thermal subsidence (WALLEY, 2000) and fault reactivation, for example in the intraplate Palmyrides basin, as far as the Sinjar basin and Euphrates graben (SAWAF *et al.*, 2000). The Erathosten - Rhodos - Bay Daglari isolated Tethyan block, which lie not far from the west, may be too the source of clastic input within the Levantine Jurassic sediments (HIRSCH *et al.*, 1995); therefore, the solution adopted on the maps is a deep but narrow subsiding furrow, west to the present day Levantine coast line but without oceanic crust during the whole Jurassic.

Following or not the hypothesis of a crustal separation and the development of a NNE-SSW trending Levantine margin, a Late Triassic - Liassic phase of extension settles a complex of uplifted or subsident blocks: the Negev and Galilee highs in Israel, the Lebanon and Coastal Chain in Lebanon and Syria, the Hamad uplift in Syria and Jordan, the Aleppo - Mardin plateau, Rutbah high in Syria - Turkey – Iraq, the Sinai deep in Eastern Egypt, the Judean embayment in Israel, the Anti-Lebanon in Lebanon, the intraplate Palmyrides rifting basin, the Sinjar depression and Euphrates graben in Syria. During the whole Jurassic these areas play as emerged areas or coastal shallow to deeper marine platforms and basins.

The Late Sinemurian interval falls in a period dominated by emersion of the Levant. The boundary between the Triassic and Jurassic strata shows a more or less clear disconformity, in several places underlined by alteration-erosion and gaps, prior to marine clastics - carbonated facies.

In Southern Israel, Negev High and northern Sinai (clastic - carbonatic area), a bauxitic - lateritic palaeosoil with flints and clays, is considered as the oldest Jurassic layer. Up to a regional erosional - karstic surface, it lies unconformably either on the Norian - Rhaetian or Early Liassic dolomitic and sandy facies (HIRSCH *et al.*, 1998). In Northern Israel and Lebanon - Galilee High (carbonatic area), it is partly replaced by the Asher volcanics, which age is bracketed 180-200 Ma (KOHN *et al.*, 1993). Contrary to what was previously claimed (Pliensbachian; PICARD & HIRSCH, 1987), a younger age for the Jurassic transgression (latest Sinemurian - earliest Carixian) is very probable as indicated by dinoflagellate cysts and larger foraminiferas. Hence, platform carbonates has been reported in continuity from the Sinai deep to the Central Israel Judean embayment while the back of the latter is filled with continental facies.

The oldest Jurassic sediment in Lebanon is a limy and dolomitic facies with laminites and collapse breccias, deposited in a shallow water peritidal environment. Shales and clastics interfinger with limestones in south Anti-Lebanon. The time equivalence is still imprecise; Early Liassic is ascertained but Sinemurian age is conjectural (WALLEY, 2000).

In the Coastal Range and Palmyrides in Syria (MOUTY, 1997a and b; MOUTY, 2000; SAWAF *et al.*, 2000), marly to shaly facies and dolomitic limestones, deposited in restricted shallow marine environments, are certainly diachronic, upgrading in the Jurassic during the transgression on the entire region with the exception of the NE-SW elongated Hamad Uplift; they are dated Early to Late Liassic by larger foraminiferas, but without more accuracy.

Western Jordan is an emerged area in relation with the Ruthbah High - Hamad Uplift and Arabian Shield. The north Jordanian areas should be a coastline where continental to restricted and nearshore varied deposits onlap the northern flank of the Rutbah Uplift that was a relative high since at least the Early Triassic (BANDEL, 1981; ALSHARHAN & NAIRN, 1997).

The Early Liassic Asher volcanics in North Israel document a rifting episode contemporaneous with further magmatism in south Turkey, NW Syria and SW Cyprus which could be due to mantle plume activity (LAWS & WILSON, 1997). The geochemical signatures of the volcanics in Syria and elsewhere support this hypothesis. Hence, the renewed Early Jurassic rifting could be a consequence of a reactivated Levant margin which affects the Palmyrides intra-plate basin, the Sinjar basin, the Euphrates graben and the Mesopotamian foldbelt (SAWAF *et al.*, 2000). Early Liassic, as well as the entire Jurassic, does not exist in several of these regions of the north Arabian plate. Hence, similar Middle to end- and post-Jurassic uplifted events should be responsible of non-deposition or major erosion in these areas (ALSHARAN & NAIRN, 1997).

III.9.- Central Arabian platform and Gulf area - Iraqi platform - Oman - Zagros basin

A lot of new data about Jurassic has been recovered from the Arabian plate and adjacent areas. Several synthesis has been published, overall dealing with sequence stratigraphy interpretation (LE NINDRE *et al.*, 1990b; GRABOWSKI & NORTON, 1995; LE MÉTOUR *et al.*, 1995; ALSHARHAN & NAIRN, 1997; AL-HUSSEINI, 1997); than, it is recommended to refer to these documents for details about the succession of formations, the sea-level variations and the geodynamic history of the north-eastern part of west Gondwana.

The Sinemurian is elsewhere missing in Arabian peninsula; it is demonstrated that a gap follows the end Triassic - Early Liassic regression which emphasises the major unconformity at the top of the Late Triassic. The Arabian platform, including the Dhofar High and Oman, and the Nubian platform are totally emerged. However, Sinemurian may exist somewhere in the Gulf area (ALSHARAN & NAIRN, 1997; GRABOWSKI & NORTON, 1995) or in the subsurface onshore of Kuwait (CARMAN, 1996; YOUSSIF & NOUMAN, 1997) and western and southern Gulf (AL HUSSEINI, 1997).

In Oman, emerged areas show rubefied weathered rock with siltstones, fine to coarse grained sandstones and ferruginous shales of continental facies, may be Early Liassic deposits (Hettangian - Sinemurian); they overlie shallow-marine Norian - Rhaetian carbonates (BÉCHENEC *et al.*, 1993; LE MÉTOUR *et al.*, 1995) and they are overlain by Middle to Late Liassic marly - carbonated deposits (Pliensbachian - Toarcian). Despite the lowering of mean sea level, the Musandam Peninsula is the only region of the Arabian plate, south of the Zagros basin (ALSHARAN & NAIRN, 1997), which is still submerged at the end of Triassic and beginning of Jurassic; limestones and shaly-limestones are interpreted as Sinemurian. The continued shallow to deeper marine sedimentation illustrates the active subsidence affecting this region since the Late Permian.

In Iraqi platform, various sandy - silty to clayey formations, carbonated and dolomitic or evaporitic, of shallow marine to coastal plain and fluvial environments are reported as Early Jurassic (ALSHARHAN & NAIRN, 1997).

III.10.- Ethiopia - Somalia - Gulf of Aden - Yemen

The oldest Jurassic deposits of the "African Horn" are "pre-Toarcian" continental to marginal marine "molassic-like" siliciclastic facies which unconformably lie on the Pan-African basement or the Karroo sediments (BOSELLINI, 1989). The age of these fluvial to coastal plain siltstones-sandstones varies from Triassic (Blue Nile basin) to Liassic (Mandera - Lugh and Al Madoh basins) and Middle Jurassic (Ogaden, Mudugh, Berbera and Borama basins); therefore, it would probably include the Sinemurian. In the Ahl Mado region, the Early Liassic age is deduced from the conformably overlying marine Early Toarcian (ABBATE *et al.*, 1974; BOSELLINI, 1989; LUGER *et al.*, 1994a).

The large varying thickness of the widely distributed deposits in Somalia and Ethiopia is in relation with NE-SW and NW-SE rift structures which first time activation is Pre-Jurassic to Early - Middle Jurassic. Fluvial and coastal plain to marginal marine sedimentary figures (braided rivers and low sinuosity meandering rivers with Coniferophyte wood pieces) witness the erosion and the transport of sediment from feebly uplifted areas to depressed basins, in tropical climatic conditions (BEAUCHAMP & LEMOIGNE, 1972, 1974; BEAUCHAMP, 1978; LUGER *et al.*, 1990).

Although the sea certainly flooded the region in the Pliensbachian, the beginning of the transgression would take place in Late Sinemurian (BOSELLINI, 1989) at the onset of the rifting which is emphasised in the Middle Jurassic prior to the break-up of the "African Horn" from the Indian - Madagascar plate in lattermost Jurassic - Early Cretaceous.

At the beginning of the Jurassic N-E trending basins exist in Yemen and Northern Somalia. They are filled with fluvio-lacustrine shales and sandstones (BEYDOUN, 1989, 1997; BEYDOUN *et al.*, 1996; AL THOUR, 1997) which age should be Liassic; subsequently, they are reported both on the Sinemurian and Toarcian maps.

East Africa seams to have no marine connection with the Tethys realm, neither through the so-called "Pakistan portal", nor with the Central Arabian platform which is a huge emerged area.

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8.- MIDDLE TOARCIAN (180 - 178 Ma)

J. THIERRY[1]

I.- MAIN FEATURES

The Middle Toarcian map was already present in the Tethys programme (BASSOULLET *et al.*, 1993); it illustrates the palaeogeography during the last stage of the Liassic subsystem. The subdivision of the Toarcian into three parts is internationally admitted in order to make easier the correlations between NW Europe and west Tethyan domains. Within the Middle Toarcian, the boundary between the Bifrons and the overlaying Variabilis Zones generally corresponds to a strong sedimentary alteration, associated with a faunal turnover which was previously used to distinguish an Early and a Late Toarcian.

The selected time slice is the Bifrons Zone, henceforth considered as the base of the Middle Toarcian (ELMI *et al.*, 1997); the overlaying Variabilis Zone may be included if it has the same facies in some places of NW Europe. On the contrary, the underlaying Serpentinus Zone (NW Europe realm) or Levisoni Zone (Mediterranean realm) are generally excluded because they are more often represented by easily distinguishable facies: condensed with ferrugineous oolites and/or manganiferous or phosphate crusts, "Schistes carton", "Paper Shales" and "Posidonien Schifer" (JENKYNS, 1984, 1988). Where endemic ammonite faunas exist, for instance on the Arabian platform, the "Nejdia faunas" of the Brankampi Zone are a good equivalent for this time interval (ENAY *et al.*, 1987; ENAY & MANGOLD, 1994).

In spite of a provincialism which progressively settles only up to the Bifrons Zone, the enough homogeneous ammonite faunas of the lower and Early - Middle Toarcian allow to use a common zonal scheme over the majority of the Peri-Tethys areas. Very good correlations are possible on all the NW European and the Mediterranean realms, as far as Moesia, Crimea and Caucasus (KALACHEVA, 1988). Sligh differences in ammonite associations on the southern areas allow however good datings and correlations all over Maghreb, as far as on Arabian platform (ENAY *et al.*, 1987; ENAY & MANGOLD, 1994) from where endemic taxa migrate westwards onto Morocco (GUEX, 1973).

In areas with scarce or without ammonites (West, East, Central and SE European platforms, Maghreb), brachiopods are diversified (ALMERAS *et al.*, 1997); somewhere collected with ammonites in several distinct facies and palaeoenvironments (fine grained to shaly - clayish coastal or shallow marine, platform carbonates and deeper carbonate - hemipelagic oozes), they allow direct correlations and a good enough biochronostratigraphic resolution at the substage accuracy.

Still scarce and not diversified, but in various deposits, calcareous nannofossils (DE KAENEL *et al.*, 1996; GARDIN, 1997), dinoflagellate cysts (RIDING & IOANNIDES, 1996; FAUCONNIER, 1997), ostracods (BODERGAT, 1997) and smaller benthic foraminiferas (RUGET & NICOLLIN, 1997) can be correlated with ammonites, mainly in areas with silty-clayey carbonaceous facies. The time accuracy is the substage or sometimes the ammonite zone.

Late Liassic or Toarcian may be identified with spores and pollen in continental deposits of the African craton or marginal shallow environments of European craton; with very rare vertebrate rests they are therefore of palaeoenvironmental interest.

Because of the lack of magnetic polarity data from oceanic crust, several time scales exist for the Toarcian but a precise and global scale is not yet accepted. Within the composite available scale (OGG, 1995; GRADSTEIN *et al.*, 1995), the Toarcian magnetic polarity pattern is compiled from a revised stratigraphic correlation of several studies (GALBRUN *et al.*, 1988, 1990), including data from the Thouars stratotype, southern Switzerland sections and *ammonotico rosso* formation in Epirus (GALBRUN *et al.*, 1992, 1994). Considering that ammonite zones are calibrated within the stages and are assumed to have equal duration, the Bifrons Zone coincide with two successive normal and reverse magnetochrons near 185-187 Ma. The palinspastic reconstruction used is that of the previous Tethys map (BASSOULLET *et al.*, 1993; RICOU, 1996), based on the palaeomagnetic data linked to the

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In conclusion, the Middle Toarcian map is assumed to illustrate the Peri-Tethyan palaeogeography near 180-178 Ma. This proposal is consistent with the only available nearest radiometric data which bracket the Toarcian: the older is in the Pliensbachian, near 194.1 ± 0.6 Ma (THOMSON & SMITH, 1992) at the boundary between the Ibex/Jamesoni Zones (Early to Middle Carixian); the younger is in the Early Bajocian near 173.5 ± 2.6 Ma (ODIN *et al.*, 1993), at the boundary of the Discites/Laeviuscula Zones.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Laurasia and West Gondwana are the major plates concerned. The kinematics trends remain similar to the Sinemurian in the context of the moving of Laurasia and Gondwana with the respectively south and north shifting of their eastern and western parts:

1.- the evolution of the Peri-Tethyan regions from the Trans-Caspian areas onto North Dobrogea are both controlled by the subduction of the Tethys oceanic crust and the opening of the Crimean - Caucasus furrow marked by active extension;

2.- the Pangaea is still in a relatively slow brake-up phase except the Central Atlantic which proceeds a rifting activity culmination (STEINER *et al.*, 1998; ZIEGLER *et al.*, 2000) keeping up pull apart basins along the present day West Morocco coast. Active extension still plays in several areas between West European and Saharan cratons. Several furrows and basins, previously installed in Ligurian surroundings, lengthen in N-NE and S-SE directions, but not yet with oceanic crust in their proximal parts. Passive thermal subsidence still plays in the Polish Trough and several NW European basins;

3.- the western propagation of Tethyan branches is emphasised by their widening and lengthening: the Vardar - Transylvanian and Bükk oceanic furrows isolate the Ticza block from Moesia and SW - Central Tethyan platforms. Several connected deep sea areas dissect the margin of Europe which still appears as a complex patchwork of blocks, more or less clearly isolated. The long and deep Pamphylian basin, still separates the Arabian promontory from the SW Central Tethyan platforms;

4.- the border of the Iran - Arabia craton is still a large continental passive margin facing the ocean and widely opened to the east. During the Toarcian, the rifting between Arabia - Somalia initiates, producing marine connections between Tethyan areas and Indian - Malagasy basins, through Somalia and Ethiopia platform.

A set of main blocks can de delimited in the Middle Toarcian time:

1.- the East European platform is ever part of the Laurasia. The small Kharkov basin, inserted between the Ukrainian shield and the Russian platform, is partly invaded by marine influences through several possible ways on North Crimea and Azov - Kuban depression;

2.- the Precaspian areas and the Turan plate show a vast northwards extension of a complex continental basin which opens southwards on the Great Caucasus marine basin;

3.- the wide basin which extend from North Dobrogea to Caucasus, through South Crimea, still marked by island arc volcanics now located in Eastern and Western Pontides mountains, still receives an abundant detritic infilling, related to the huge north emerged areas;

4.- the Iranian block, previously attached to the Turan plate is now separated by an E-W oriented succession of troughs (Great Caucasus and Lesser Caucasus) and ridges (North Trans-Caucasus);

5.- the Moesian block is underlined by the North Tethys subduction zone;

6.- the succession of troughs and basins that border the East European - Scythian platform and the Fenno-Scandian shield, from Dobrogea to the Ergesund - Danish basin, through the East Carpathian Gate, Baltic platform and the Polish trough, are below marine conditions. The latter shows a discreet episode of accelerated subsidence; it separates the huge north shields from the emerged Bohemian massif;

7.- the Mid, West and North European areas keep to enlarge their connections in every direction, as a consequence of constant marine invasions on lowlands; they still more appear as an archipelago of platforms or more or less emerged areas separated by large basins and troughs, inserted between the Fenno-Scandian and Greenland shields;

8.- the eastern façade of the Iberian block shows unchanged connections with West Europe along the tectonic alignments of the future Bay of Biscay rift. Its Atlantic (Lusitanian basin) and south (Subbettic domain) rims proceed with a rifting regime; communications with Boreal realm, Maghreb areas and the originating Atlantic domain are reinforced along the so-called "Hispanic Corridor";

9.- from Morocco to Tunisia, the Atlasic domain is still below intense rifting conditions with active extension; asymmetric basins open on the Atlantic side of Morocco;

10. - the Saharan areas and connected continental basins still register active extension;

11.- the present day shoreline of Libya and Egypt remains in passive margin conditions; continental basins, endoreic or in connection with the marine domain, still extend on the Libyan - Egyptian - Nubian part of the African craton;

12.- the Levant and Arabia - Iran areas are still passive continental margins with asymmetric width of their shelf parts. As a consequence of the world-wide Toarcian transgression, the latter is mostly invaded onto the mid part of the vast Arabian block, while east embayments grow up on the former. A marine connection is still assumed between the two shelves by mean of an E-W platform below fluctuating shallow marine conditions;

13.- the complex system of grabens and highs of the Somalian - Ethiopian areas is drowned by marine waters, mainly coming directly from Tethys, driving to an extension of a large shelf, far to the west on the Nubian craton.

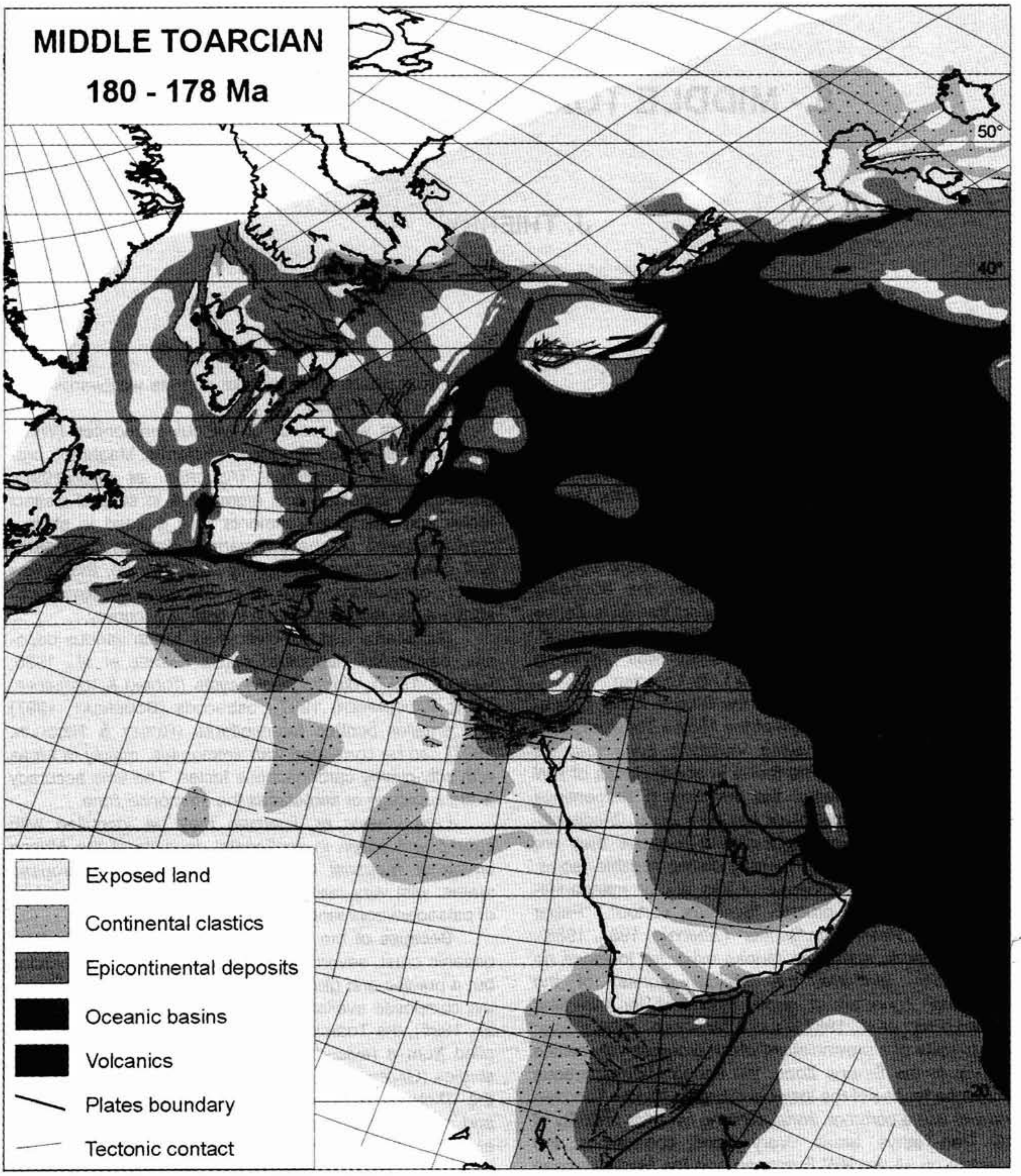

Fig. 8.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Middle Toarcian.

so called "East Coast Anomaly" (KLITGORD & SCHOUTEN, 1986); this anomaly fairly fits to the selected Middle Toarcian chronostratigraphic interval.

No radiometric data is directly correlated with the Toarcian ammonite zonal scheme. According to the ODIN's time scale (1994), the numerical ages for the whole Toarcian fall in the interval of 175-184 Ma. Using an equal duration for each ammonite zone, the time slice concerned for the Bifrons Zone should be between 180-178 Ma. Such age is slightly younger than that deduced from the GRADSTEIN's time scale (1995), 185-187 Ma, whose Toarcian lower boundary is near 189 ± 4 Ma. and the upper boundary at 180 ± 4 Ma.

II.2.- Palaeoposition of plates and blocks

Alike for the Sinemurian map and because no available new data exist, the kinematic setting of the plates is that of the Toarcian Tethys map (RICOU, 1996); so, there is no computed new position of blocks and the palaeolatitude grid is unchanged (BESSE & COURTILLOT, 1991). To be in harmony on all the successive maps, the position of Iberia and Corsica - Sardinia blocks are modified if compared with the Tethys Programme maps (FOURCADE *et al.*, 1996; OLIVET, 1996; VIALLY & TRÉMOLIÈRES, 1996). According to its position on the Sinemurian map, Moesia is slightly eastward shifted and rotated in order to illustrate the Magura - Ligurian rifting and the opening of the Vardar oceanic trough. Such a position is in agreement with the palaeolatitudes data of Moesia and the Eurasian palaeolatitude curve (SURMONT *et al.*, 1991; TCHOUMATCHENCO *et al.*, 1992).

II.3.- Accuracy

The magnetic sequence with numerous inversions compiled from the Toarcian stratotype in the Paris basin and in sections from South Switzerland, does not yet provide a synthetic and reliable scale, even though some data are calibrated on ammonites zonal scheme. The palinspastic reconstruction is that of the classical Atlantic fit.

The biostratigraphic data allow very good correlations within a high resolution frame which depends on the fossil group (ammonites, brachiopods, etc.). Overall in NW Europe (RIOULT *et al.*, 1991; GRACIANSKY *et al.*, 1998a and b; STEPHEN & DAVIES, 1998; HESSELBO & JENKINS, 1998; DUMONT, 1998), sequence stratigraphy provides very good correlations when they are constrained by biostratigraphic data. The detailed succession and the hierarchy of the several order sequences and cycles compiled on the Jurassic sequence Chronostratigraphy/Biochronostratigraphy Chart (HARDENBOL *et al.*, 1998) is henceforth used as a reliable framework for stratigraphic correlations.

II.4.- General comments

The Middle Toarcian sedimentary features coincide with a period of major subsidence, related to continuous rifting events and active extension tectonics quite all over the Peri-Tethyan areas, along passive margins or in grabens under marine or continental conditions (North Sea, Polish Trough, North Dobrogea, Crimea, Caucasus, Paris basin, Sub-Alpine basin, Atlasic and Tunisian Trough, Chott basin, East African Corner basins, etc). Since the Sinemurian and during the Pliensbachian, the rifting activity had reached a noticeable intensity which will go on till the lower Callovian, putting at the foreground a major trend of the "Ligurian Cycle" (JACQUIN & GRACIANSKY, 1998).

At a time of important and world-wide scale transgressions, the Middle Toarcian map is established at the period that immediately follows or coincide with a high sea level which corresponds to a major peak transgression dated as around the Early to Middle Toarcian boundary (Falciferum/Sublevisoni subzones or Serpentinum/ Bifrons Zones boundaries; HARDENBOL *et al.*, 1998). The Toarcian peak transgression is characterised by sediment starvation on very shallow platforms, coeval with the development of silty-clayey carbonaceous organic rich shales in offshore environments. The anoxic event is more marked on the Laurasian epicratonic seas than in the Gondwanan but it is considered to have a world-wide extension (BAUDIN & HERBIN, 1996); the resulting deposits are the main kerogenous source rock of Western Europe. Black shales and thick source rocks occurred in the European terrigenous shelf, whereas coeval thin horizons exist in the Mediterranean Tethys. Coal bearing interbedded layers occur elsewhere in eastern Tethyan realm and locally in western Gondwana, witness to the situation of the considered areas in the humid climatic belts, either of the tropical zone or the equatorial. Following the peak transgression, the thick marine sedimentary sequence registered a transition from calcareous to argillaceous deposits, the latter becoming rapidly dominant. In epicontinental platforms as well as in deeper basins, abundant necto-pelagic faunas (ammonites and belemnites) are a striking feature.

During the Toarcian, radiolarites are restricted to several specific places of the Tethys domain as it was in the Sinemurian (DE WEVER *et al.*, 1996; FOURCADE *et al.*, 1996) while nodular *ammonitico rosso* type and pelagic limestones are particularly well developed and spread everywhere (CECCA *et al.*, 1992).

Keeping on a sea-level rise tendency since the Early Liassic, the generalised transgression of the Toarcian continues to enlarge the Peri-Tethyan shelves. Most of the marginal platforms and basins are concerned, everywhere on the Gondwanan plate, but evaporitic or brackish deposits are still largely extended. The importance of the shallow carbonate deposition decreased probably due to a reduction in productivity, a general deepening induced by the transgression and the increase of terrigenous input. Carbonate platforms are developed only in few places of the northern Tethyan areas, but they widely extend in west central Tethys and on its southern margin. Coastal and shallow marine terrigenous facies still underline the emerged north continent while deep carbonates and pelagic facies spread on its shelves.

The facies of the Toarcian sediments show significant and classical latitudinal distribution which fairly fits with the palinspastic reconstruction and grossly shows significant features which allow climatic interpretations; but, the sea-level changes seem to be more strongly related to local and regional tectonic events than to global factors. The evolution from carbonaceous to shaley sedimentation, observed from Hettangian to Toarcian, suggests that the long-term Liassic transgression and climatic evolution, reputed to be warmer than the Present and in more equable conditions at high latitudes, would have no strict causal link between each other (JACQUIN & GRACIANSKY, 1998).

Marine connections are proved between the Boreal domain and NW Europe through the Greenland - Scandinavia seaway (CARIOU *et al.*, 1985; DORÉ, 1991, 1992). Several weak north marine ingressions exist on the southern border or the Russian platform, from Crimea

- Caucasus basin, but this huge cratonic area remains quite totally emerged. It is no more possible to confirm communications between the European - Maghrebian areas and the Western Carribean Tethys through the Atlantic areas and the Hispanic seaway (ENAY, 1980; CARIOU *et al.*, 1985; BASSOULLET *et al.*, 1993; FOURCADE *et al.*, 1996).

III.- DEFINITION OF DOMAINS

III.1.- Russian platform: Volga - Ural - Donetz - Ukraine

The Russian platform is still a large emerged area which undergoes strong erosion; it is connected with the Fenno-Scandian shield, the Ural highs, the Ukrainian - Stavropol shields and the western Precaspian high (VINOGRADOV, 1968). Coastal-shallow marine detrital deposits, certainly related both to a passive thermal subsidence without conspicuous faulting (STOVBA *et al.*, 1996; VAN WEES *et al.*, 1996; STEPHENSON *et al.*, 2000), and to the Middle (Pliensbachian) and Late (Toarcian) Liassic transgressions, invade the Kharkov - Donetz basin. This latter one is west bordered by continental clastics and in connection with the North Crimea - Azov rifts, and the Kuban fore-deep/depression - Terek basin. However, the present day limits of the Jurassic sequence are erosional; the original depositional area of the Donetz area would be wider (ULMISHEK *et al.*, 1994).

III.2.- Turan plate

Intracontinental (limnic), fluvial and coastal (paralic) deposits, supposed to be Toarcian (or older), fill the depressions, bottom and mouth of palaeovalleys in North Ustiurt - South Mangyshlak and Trans-Caspian areas. Extending to the north into the Aral basin, onto the Precaspian areas, the distribution of the sediments may be once more regarded with the alternative sub-latitudinal or sub-longitudinal orientations (PANOV *et al.*, 1996; VOLOZH *et al.*, 1997). South and east of the emerged Kara Bogaz high, in the Greater Balkan and South Karakum platform, sandstones with argillites, sometimes coal bearing (Daghestan), with thin bauxites-like layers or interbedded conglomerates, fill a tilted block framework (LYBERIS *et al.*, 1998; LYBERIS & MANBY, 1999). In Kopet Dagh, a thick sedimentation of fluviatile to coastal - shallow water siliciclastic environments, install along a margin which is the eastern prolongation of the Caucasus - South Caspian trough; deposits are dated Middle Jurassic by palynomorphs, but they certainly include the Late Liassic. The central Karakum, Kara Bogaz and several other denudation areas situated further to the east provide the terrigenous input. The tectonics and basin development is still a NE-SW extensional setting, accommodated by a normal faults trending (THOMAS *et al.*, 1999).

III.3.- Scythian platform - Crimea - Black sea - Caucasus and Precaspian areas

The Trans-Caucasus - Pontides N-dipping subductional magmatic belt which had originated since the Early Liassic, generates a major back-arc rifting with deep-water basins along the South Caspian - Great Caucasus - Scythian platform - Black Sea - South Crimea - Dobrogea belt (ROBINSON *et al.*, 1996; ROBINSON & KERUSOV, 1997; NIKISHIN *et al.*, 1998a and b, 2000). The Precaspian areas remain continental.

In Pliensbachian - Toarcian, extensional movements and cooling of the lithosphere (ERSHOV *et al.*, submitted) continue to develop the Kuban and Terek basins in N Caucasus, while the Stavropol - Voronezh high remain a stable area, as it was before; a strong inheritance seems to be the character of this structure all over the Jurassic. The Pliensbachian - Toarcian sediments (mainly sandstones, siltstones and claystones of coastal to shallow marine environments) are enough difficult to separate in spite of ammonites and foraminiferas biostratigraphic data in central and west Caucasus (PANOV *et al.*, 1994). Similar facies are located in narrow rift-like troughs in North Crimea (Karkinitsky - Azov basins), Scythian platform (Kuban basin) and north Caucasus rim (Terek basin), which are progressively drowned during the two successive major Middle and Late Liassic transgressions, tectonically enhanced by the rifting phases.

The weak rifting phase of the previously uplifted Scythian platform contrasts with the stronger setting of the Great Caucasus deep-water rifted trough (Late Pliensbachian - Toarcian - Early Aalenian; NIKISHIN *et al.*, 1998a and b, 2000), where crustal thinning (or local spreading) and dacite - andesite - basalt volcanism is inferred to occur in Toarcian (PANOV & GUSHIN, 1987). This major phase of back-arc rifting which controls the Great Caucasus - Pontides tectonic and sedimentary history is certainly responsible for the Dobrogea and Stara - Planina evolution in the Moesian block. South Caspian areas are not well stratigraphically constrained, but they would undergo the same evolution as the Great Caucasus. The Iranian plate is progressively drowned (DAVOUDZADEH & SCHMIDT, 1983).

The continental - fluvial terrigenous series of the northern slope of the Great Caucasus and Precaspian areas are assumed to grade into the Toarcian. Much more developed, they are reported following the intermediate solution already discussed for the Late Sinemurian map, referring to a sub-latitudinal (PANOV *et al.*, 1996) or a sub-longitudinal (VOLOZH *et al.*, 1997) orientation of the fluvial system. Whatever would be the orientation, these fluvial systems strongly feed the Black Sea - Caucasus belt sedimentation with varied clastics input.

III.4.- Teisseyre/Tornquist zone - Moesian platform

The Polish Trough palaeogeography during the Toarcian is not very different from that of the Sinemurian

and Pliensbachian, but the surface extent of the basin progressively decreases. The basins flanks are slightly uplifted with a coeval increasing of both sediments thickness and subsidence rate; however, the latter does not reach great values (LAMARCHE *et al.*, 1998). Up to its western prolongation along the Teisseyre - Tornquist zone in the Ergesund - Danish basin, it is still north bordered by the uplifted Fenno-Scandian - Ukrainian shields and Scythian platform, the emerged areas supplying clastics to the basins. Its southern border, the Bohemian massif, a permanent land mass during the whole Jurassic (MALKOVSKY, 1987; ZIMMER & WESSELY, 1996), similarly shed clastics into the adjacent areas. Elsewhere, the north marginal zones are dominated by fluvial to deltaic - limnic and shallow marine terrigenous sedimentation, where the biostratigraphic control is given by megaspores and sporomorphs. On the Baltic platform (MAREK & GRIGELIS, 1998), mudstones and sandstones are accumulated in a shallow-water brackish bay which is gradually invaded by sandy sediments transported by rivers flowing from the NE.

In Early - Middle Toarcian, marine ingressions covered large areas of the Polish lowlands, as far as the Baltic platform and the Bohemian massif rim. However, in contrast to the Pliensbachian, these ingressions did not induce the formation of open sea sediments. The conditions are equalised over large areas where monotonous siltstones and fine grained sandstones accumulate in shallow water environments; the biostratigraphic control is given by agglutinated foraminiferas and bivalves (PIENKOWSKY, 1997; DADLEZ *et al.*, 1998; MAREK & GRIGELIS, 1998); depositional sequences and transgressive - regressive cycles are described and correlated with the Western European chart (FELDMAN-OLSZEWSKA, 1997a). The Danish - Ergesund basin and Mid Polish trough shallow marine sandy-clayey facies are connected with the monotonous North Germany - Hannover epicontinental open sea and shaly deposits, the so-called "mixed facies" and "non-source rocks" of the "green series", referring to their organic rich potential. These claystone facies show a gradual eastward decrease of the source-rock potential of the Lower Toarcian Posidonia Shales.

The geometry, facies and palaeoenvironments deposits of the Polish basin, indicate rifting conditions with a uniform and weak subsidence pattern guided by the activity of synsedimentary bounding faults along a transverse asymmetric half graben basin NE dipping; intrabasinal uplifts (Wielkopolska High) are several times active (DADLEZ *et al.*, 1995; LAMARCHE *et al.*, 1998; LAMARCHE, 1999; STEPHENSON *et al.*, submitted). The basins install in a N-NE - S-SW transtensive and E-SE - W-NW transpressive regime, with a generalised senestral component along the Teisseyre - Tornquist zone (HAKENBERG & SWIDROWSKA, 1997).

More eastward, in the Ukrainian Carpathians areas, active subsidence along the margin of the Scythian platform drives to a fluvial - alluvial to deltaic sedimentation, which grades into limnic coastal marine to terrigenous shallow marine. The latter are related to the eastward flysch-like Toarcian deposits of the North Dobrogea (GRADINARU, 1993). During the Toarcian, most of the foreland was an emerged area, except for North Dobrogea and the western part of the Pre-Dobrogean depression where quartzose sandstones and sandy clays appear (MOROZ *et al.*, 1997). Continued differential uplift and subsidence is reflected in the propagation of E-W big faults (Galati – Sf. Georghe, Peceneaga Kamena, Capidava - Ovidiu, Intramoesian; GRADINARU, 1984, 1988) which delimitate several north to south successions of tectonic units (North, Central and South Dobrogea). Coarse clastics are deposited near the active front, while turbiditic fans prograde to the north and east, infilling the faults delimited basins (BANKS, 1997; BANKS & ROBINSON, 1997; TARI *et al.*, 1997; SEGHEDI, 2000).

On the Bulgarian side of the Moesian platform, the Stara Planina basin rift-south facing passive margin extended eastwards: south thickening of the series, normal faults and gradual subsidence (BANKS, 1997; TARI *et al.*, 1997) are probably a response to crustal extension (NIKISHIN *et al.*, 2000). The principal fault sets (NW-SE, NE-SW, E-W) and horst - graben system are inherited from the earliest Jurassic (TCHOUMATCHENCO *et al.*, 1989; TCHOUMATCHENCO & SAPUNOV, 1994). The sea enlarged its area and gradually begun to occupy the peripheral parts of some previously uplifted areas (Vidin, Vraca, Pleven and East Balkans). The existing grabens (Sevlievo, Mihajlovgrad, Izdremets, Veliko - Tarnovo, Provadija) and furrows (Nis-Trojan and Trecjlano troughs) are still active (SAPUNOV & TCHOUMATCHENCO, 1990). New grabens are created (Breznik and Svetlja grabens), witness of the continuation of the E-W intra-plate disintegration initiated in the Sinemurian and Pliensbachian. The marine deposits are dated by ammonites, brachiopods, small foraminifers and spores and pollen. The sedimentation is mainly terrigenous and calcareous, predominantly characteristic of inner shelf to restricted environments, sometimes of outer shelf (SAPUNOV & TCHOUMATCHENCO, 1987; SAPUNOV *et al.*, 1988, 1990, 1991; TCHOUMATCHENCO & CERNJAVSKA, 1989). The Thracian - Rhodopes massif is still emerged, east interrupted by the shallow marine Strandzha area.

III.5.- Western Europe platform

The Toarcian palaeogeographic evolution of the W European platform is driven by the "Ligurian cycle" (GRACIANSKY *et al.*, 1998 a) which sea level variations are related to the Ligurian rifting (LEMOINE & GRACIANSKY, 1988; DUMONT, 1998) and the opening of North Atlantic. Commonly known in all west European basins and immediately following the peak transgression dated Falciferum/Serpentinus Zone (latest Early Toarcian), the regressive phase begins and spans over the Middle/Late Toarcian to Late Aalenian. Contrasting with the peak transgression sediment starvation and the development of carbonaceous shales in offshore environments, the early steps of this regressive phase are characterised by shaley sediments which largely dominate sands and carbonates in all latitudes.

The Toarcian sea-level rise leads to the drowning of the majority of the uplifted areas and the overlaping of the basin margins driving to enlarge the epicontinental sea. The present day extension of the Toarcian deposits should be previously greater: few marginal environments are registered around the emerged areas; they should be

reduced or later eroded (Toarcian - Aalenian boundary) at time of the "Mid-Cimmerian unconformity", as a consequence of the North Sea thermal doming (UNDERHILL & PARTINGTON, 1993; JACQUIN & GRACIANSKY, 1998). Several basins easily communicate each other as a consequence of reduced emerged areas, some of which (Ebro massif in Spain, Massif Central in France) should be totally submerged. The phases of the transgression are very well documented; they are divided into high resolution and very well dated depositional sequences which can be correlated quite all over the Western Europe (GRACIANSKY *et al.*, 1993, 1998a and b; HESSELBO & JENKYNS, 1998; STEPHEN & DAVIES, 1998).

The sedimentation is everywhere dominated by clays and silts, in places interbedded at the base with organic rich shales in North Germany - Hanover basin, south North Sea, Paris basin and several U.K. basins (FLEET *et al.*, 1987; CAMERON *et al.*, 1992; HOWARTH, 1992; DAMTOPF *et al.*, 1992; HAMBLIN *et al.*, 1992; BINOT *et al.*, 1993; MORTON, 1987, 1989, 1990, 1993; BESSEREAU & GUILLOCHEAU, 1994; HANZO & ESPITALIÉ, 1994; VAN ADRICHEM-BOOGAERT & KOUWE, 1997; WIGNALL & HALLAM, 1991). Because of diachronic deposition of the shales and some doubtful datings in subsurface data, the time slice and facies package reported on the map for the West European area may be comprehensive. They will partly encompass the organic rich sediments prior to the peak transgression ("Schistes carton", "Paper shales" and "Posidonyen Schiffer"), the condensed beds (ferruginous oolites, manganiferous or phosphatic crusts and nodules) at the maximum transgression, and the shaley beds which immediately follow the peak transgression: "Drake Formation" in the North Sea (STEEL, 1993), "Marnes supérieures" and "Marnes à Bifrons" in the Paris basin, "Alum Shales Formation" in the Yorkshire basin (POWELL, 1994), upper part of the Schwabian "Brauner Jura", "Marnes de la Javie" in the Digne sub-Alpine basin (GRACIANSKY *et al.*, 1993). For all these areas, the facies recorded on the Toarcian map bracket the classic more or less bituminous - organic rich sediments of black-shales ("source-rock"), deposited on medium deep distal platform environments, and the lateral or overlying "mixed facies"; the latest Early Toarcian to earliest Middle Toarcian age is based on microfossils and palynomorphs in subsurface, on ammonites in outcrops. As examples, in the West Netherlands (VAN ADRICHEM-BOOGAERT & KOUWE, 1997) and adjacent parts of Germany (BRAND & HOFFMANN, 1963; WEITSCHAT & HOFFMANN, 1984; KÖLBEL, 1968; BOIGK, 1981; KOCKEL, 1995), the "Posidonia Shales" extend up into the Bifrons Zone, and can be locally somewhat younger. In the Sole Pit basin (U.K.), there is indication that organic shale deposition did not extend so far up in the Late Toarcian. In the Digne - Subalpine basin (France) the condensed level (peak transgression) may reach the Bifrons Zone (GRACIANSKY *et al.*, 1993).

Sandy facies underline some NW European remnant emerged areas (Shetland platform, Armorican and Bohemian massifs). The carbonate platforms are restricted to the rim of few SW European emerged massifs (Corsica - Sardinia - Provence, Iberian, Lusitanian platforms).

In the future Alpine areas (Ardèche, Dauphinois basin, Valais trough, Briançonnais) the tectonic activity is reduced to extensional movements; a tilted blocks framework delimits ridges and basins (ELMI, 1990; DUMONT, 1998). The clayey sedimentation is controlled by the effects of the subsidence and the rifting activity.

On the Iberian peninsula, in contrast with the evolution of the Lusitanian and the Betic margins which are continued rift systems, the development of the Iberian platform is governed by a regional thermal subsidence. In the Lusitanian basin, the eustatic variations are complicated by a strong acceleration of a differential subsidence which spans from Pliensbachian to Early Aalenian (SOARES *et al.*, 1993; CANÉROT *et al.*, 1995), driving to an east to west succession of facies and palaeoenvironments, from shallow platform carbonates to deeper marly successions. A main intracontinental rifting episode affects the Betic margin from Domerian onwards (VERA, 2000); the Subbetic domain is a deep water trough, whereas the Prebetic is shallow-water platform carbonates dominated. The fault activity is negligible in the Catalonian - Valencia and Iberian platforms (CANÉROT, 1989, 1991; SALAS *et al.*, 2000) which undergo a broad marly to limy sedimentation (FERNANDEZ-LOPEZ *et al.*, 1996). The Pyrenees - Bay of Biscay - Aquitanian basin evolution is more controversial; the marly to limy Late Liassic series were probably deposited during a phase of regional thermal subsidence, a quiet phase of an intracontinental rifting that separates Iberia and NW stable Europe (BRUNET, 1984; LE VOT *et al.*, 1996; YILMAZ *et al.*, 1996; VERGES & GARCIA-SENZ, 2000).

The tectonic and sedimentary evolution of the flexural intracratonic Paris basin is the overall result of a long term thermal subsidence; but, several synsedimentary tectonic trends indicate the superimposition of a short-term tectonic component due to intraplate deformation related to the several steps of the Ligurian opening (GUILLOCHEAU *et al.*, 1999; ROBIN, 1997; ROBIN *et al.*, 1996, 1998). Although there is few evidence for synsedimentary tectonics, the axis of several basins in NE Germany coincide with the trace of the Permian fault systems (MALKOWSKY, 1987; SCHRÖDER, 1987; ZIMMER & WESSELY, 1993). The Toarcian successions of the North Sea (West Netherlands, Broad Forteen, Sole Pit basins; Central graben, South Viking and North Viking graben), suggest that the sequentiality is not only caused by eustatic variations; it probably includes a major contribution from subsidence-rate variations, driven by the regional post rift-tectonics thermal subsidence (CAMERON *et al.*, 1992; PARTINGTON *et al.*, 1993a and b; MILTON, 1993; STEEL, 1993; GATLIFF *et al.*, 1994; THOMAS & COWARD, 1996; LEPERCQ & GAULIER, 1996). The thermal subsidence is responsible too of the filling of many United Kingdom inland or offshore basins (Celtic Sea, Western Approaches, Channel, Hebrides, Minches, Irish Sea, Wessex, Cleveland, East Midland; ANDREWS *et al.*, 1990; CALLOMON & COPE, 1995; COPE & RAWSON, 1992; EVANS, 1990; HAMBLIN *et al.*, 1992; HESSELBO & JENKYNS, 1995; JACKSON *et al.*, 1995; JENKYNS & SENIOR, 1991; MORTON, 1992a and b, 1993; MORTON & HUDSON, 1995; RAWSON & WRIGHT, 1995; STEPHEN *et al.*, 1993; TAPPIN *et al.*, 1994).

Favoured by the sea level rise which overlaps the Shetland platform - Faeroe high - Greenland and Norway the connections reach a maximum between the boreal and the subboreal - Mediterranean realms through the Viking Graben and the North Sea; faunal exchanges are easy across the southern North Sea, Great Britain, the Paris basin and the Franconian - Schwabian platforms.

III.6.- Maghreb (Morocco - Algeria - Tunisia) - Saharan areas

During the Toarcian transgression, the "mosaïc" episode is general over all the Atlasic domain (Elmi, 1996); the deepening of the Sinemurian - Pliensbachian furrows is more pronounced and the previously shallow marine areas (Middle Atlas and Oran Meseta; central and north Tunisian areas) are partly drowned (Peybernès, 1992; Elmi *et al.*, 1998; Poisson *et al.*, 1998; Kamoun *et al.*, 1999). The biostratigraphic record is very good in the majority of the Maghreb (Peybernès *et al.*, 1990; Bassoullet *et al.*, 1991; Soussi *et al.*, 1991a and b; Sadki, 1992; Bouchouata *et al.*, 1995; Peybernès *et al.*, 1996; Elmi *et al.*, 1998; Poisson *et al.*, 1998; Soussi *et al.*, 1998; Elmi *et al.*, 1999; Lachkar, 2000; Soussi *et al.*, 2000) with abundant ammonite faunas, brachiopods and foraminiferas, except in some areas with clastics to calcareous - dolomitic and evaporitic facies deposited in coastal plain, marginal marine and continental environments. Correlations are supported too by a consistent sequence stratigraphy framework (Souhel *et al.*, 1998).

On the Atlantic margin, the Essaouira - Agadir and Tarfaya asymmetric basins show fluvio-deltaic to marginal marine environments, sometimes with evaporitic or silty-marly facies intercalations, mainly dated by microfossils (Peybernès *et al.*, 1987; Du Dresnay, 1988; De Kaenel & Bergen, 1993; Broughton & Trépanier, 1993; Medina, 1994; Morabet *et al.*, 1998). Compared to the Sinemurian, the shoreline of the gulf located between Essaouira and Agadir displaces towards the east and the north as a result of sea-level rise and moderate thermal subsidence of the incipient Atlantic margin (Medina, 1994, 1995; Chotin *et al.*, 2000; Labbassi *et al.*, 2000). Alike during Sinemurian - Pliensbachian, the westward tilted half-grabens are the result of a NW-SE syn-rift extension (Steiner *et al.*, 1998) guided by N-NE and E-NE normal faults. No connection exits between the western termination of the Atlas trough and the Essaouira basin. The environments are rather similar while the facies are different, with clastics on the Atlantic side, exclusively carbonated (limestones and dolomitic limestones) on the Atlasic side of an enlarged and emerged Moroccan Meseta peninsula.

The peak of deepening of furrows and the coeval uplift of the shoals and furrows borders (High and Low Saharan fringe of the Moroccan and Algerian Atlas, Chotts area of the Tunisian Central Atlas) occurred during the Early Toarcian. The tectonic control of the sedimentary facies distribution is elsewhere very strong; at the end of the Liassic, the margin is segmented by normal faults individualising more or less subsident areas (Vially *et al.*, 1994; Ellouz *et al.*, submitted; Bracene *et al.*, submitted). In the Great Kabylia, the rifting which occurs in a progressive manner from the south towards the north, affects now the inner domain of the chain since the Domerian (Zizi, 1996; Cattanéo *et al.*, 1999). In the Rif basin and foreland, deltas progress from west and SW along the passive margin, from Toarcian to Bajocian (Favre *et al.*, 1991; Mehdi *et al.*, 1994; Boutakiout & Elmi, 1996; Haddaoui *et al.*, 1997). In the Atlasic troughs and furrows which fill with thick marly sediments, the rifting continues with active extension guided by an oblique normal fault regime with a sinistral transcurrent component along a NE-SW trending (Beauchamp *et al.*, 1996; Piqué *et al.*, 1998; Chotin *et al.*, 2000; Piqué *et al.*, 2000). The slopes are marked by mass-flows, calcareous turbidites and *ammonitico rosso* facies. Along the borders of the Middle and High Atlas troughs, the platforms undergo calcareous - dolomitic sedimentation which grades from marginal marine toward the emerged areas (Moroccan Meseta and Saharan craton), to shallower marine inner and outer platforms. Some of the largest shoals or shallow outer platforms develop neritic facies (Middle Atlas, Western Rhar Roubane, Tunisia); reduced and condensed ferrugineous oolitic or bioclastic layers develop on the high parts of the platform, broken up into tilted blocks since the Late Sinemurian (Elmi *et al.*, 1998; Poisson *et al.*, 1998; Soussi *et al.*, 2000).

The High Sahara is a large denudation area S to the narrow fringe of marginal marine environments. On the E, the paralic to evaporitic Lower Sahara, Oued Mya, Dahar, Ghadames and Tataouine basins receive an abundant terrigenous input ("continental intercalaire"). The large development of evaporitic deposits suggests a permanent arid climate (Lefranc & Guiraud, 1990; Busson & Cornée, 1991). Tunisia, yet in a rifting phase, is structured into horsts and grabens which undergo active submeridian extension along E-W normal faults (Barrier *et al.*, 1993; Hlaiem *et al.*, 1997; Bouaziz *et al.*, 1994, 1996a and b) and a weak but continuous and homogeneous subsidence (Patriat *et al.*, submitted); the sedimentation is evidently tectonically controlled, but climatic influences would play too (Ben Ismail, 1991). Such a regime affects the Pelagian carbonated platform until the offshore Malta (Bishop & Debono, 1996; Bouillin *et al.*, 1999; Argnani & Torelli, 2000). The Murzuq basin northward opened on the Ghadames basin and Tunisian areas, continues to fill with continental Nubian sequences of imprecise Jurassic age.

III.7.- Egypt - Sudan - Libya

The palaeogeography and the depositional framework did not change greatly since the Sinemurian despite the Toarcian is a time of important transgressions and high standing sea-level coupled with a period of widespread extensional deformation of both continental and oceanic crust (Guiraud & Bellion, 1996; Schandelmeier *et al.*, 1997; Guiraud, 1998; Guiraud & Bosworth, 1999). Extensional block faulting is evident in Sinai (Moustafa & Khalil, 1990) and north Western desert (Moustafa *et al.*, 1998; Ayyad *et al.*, 1998). As a consequence, fringed by marginal marine to intermediate environments, the marine platform carbonate deposits

shift southward all along the passive margin (GUIRAUD *et al.*, 2000).

On the Egyptian shelf, the marine transgression, which is supposed to begin in Pliensbachian, is restricted to the north Sinai, Nile Delta and extreme north Western Desert. In Sinai, tidal flat to deltaic and fluvio-marine sediments upgrade and interfinger with genuine marine shelf deposits (algal limestones and marls which yield some Toarcian ammonites (JENKINS, 1990; KERDANY & CHERIF, 1990). Deep-water environments exist north of the platform margin. A similar depositional scheme extends in the Nile Delta and the several basins of the north Western Desert (ABDEL AAL *et al.*, 1990; HANTAR, 1990; KEELEY *et al.*, 1990; KEELEY & WALLIS, 1991; KEELEY, 1994). An evaporitic facies developes laterally; it makes the southward transition to the fluvio-deltaic to fluviatile - continental "Nubian sandstones" which fill far to the south the Dakhla basin. Clastic input is still dominant. The precise dating is still unclear despite sequence stratigraphy correlations allow fairly good local correlations.

The north Eastern Desert is still an emerged area alike the Gulf of Sirt, the Cyrenaica arch until the Gargaf High and Tibesti (EL-HAWAT, 1992; EL-HAWAT *et al.*, 1996; SMITH & KARKI, 1996). A period of crustal doming and uplift is thought to characterise this part of the north Gondwanan border, while eastward, until the Levantine area, sedimentation is active in fault bounded basins.

The continental and fluviatile clastic infilling of the Dakhla, Kharga, Al Kufra, Erdis and Lakla basins is believed to encompass a Toarcian episode despite reliable biostratigraphic data are missing (KLITZSCH & WYCSISK, 1987; HERMINA, 1990; KLITZSCH, 1990; KLITZSCH & SQUYRES, 1990; KLITZSCH & SCHANDELMEIER, 1990).

III.8.- Levant (Israel - Lebanon - Syria - Jordan)

The Toarcian is represented in Israel by thick platform carbonate deposits and evaporites (HISRSH *et al.*, 1998) which onlap either laterites in Negev, or the Asher Basalts in the northern part of Israel; the latter may span until the Toarcian. It represents sediments on a slope, thickening toward central part of the Judean embayment; in Lebanon, an equivalent volcanics episode within similar facies is probable but not clear because reliable biostratigraphic data are missing. In Negev, the Early and Middle Toarcian interval may be represented by clastic facies with calcarenites which probably upgrade into the Late Toarcian - Aalenian. In the Northern Israel carbonatic Lebanon - Galilee high, the interval may be within the inferior part of a monotonous platform carbonates series. All these deposits are assumed to extend far west to the present day coast-line of Levant, probably separated from the Erathosten - Rhodos - Bay Daglari isolated Tethyan platform, by normal faults. In Southern Israel clastic - carbonatic Negev High, the Jurassic has been eroded during Early Cretaceous.

Thick massive limestones, locally with cherts and dolomitic layers in northern Anti-Lebanon area, are supposed to encompass the Toarcian - Oxfordian interval despite rare and unreliable biostratigraphic data (WALLEY, 2000). They illustrate a major drowning event (transgressive part of a low frequency Early to Middle Jurassic cycle) and the persistence of middle to outer shelf conditions; they can be partly correlated with the thick limestones of the north Israel carbonatic Lebanon - Galilee high.

The Toarcian is assumed to be within the diachronic deposits of the Coastal Range and Palmyrides in Syria (MOUTY, 1997a and b; SAWAF *et al.*, 2000), dated Early to Late Liassic by large foraminiferas but without more precision. The entire Jurassic is unknown in Euphrates graben, Sinjar trough and Mesopotamian fold belt (ALSHARAN & NAIRN, 1997) because of Early Cretaceous regional uplift and erosion.

The west Jordan is still an emerged area and North Jordan a coastline where continental to restricted and nearshore varied deposits onlap the north flank of the Rutbah High (BANDEL, 1981; ALSHARHAN & NAIRN, 1997).

III.9.- Central Arabian platform and Gulf area - Iraqi platform - Oman - Zagros basin

Several improvements has been done (LE NINDRE *et al.*, 1987, 1990b; BASSOULLET *et al.*, 1993; GRABOWSKI & NORTON, 1995; LE MÉTOUR *et al.*, 1995; ALSHARHAN & NAIRN, 1997; AL HUSSEINI, 1997) since the first published "Late Liassic" palaeoenvironments map of Arabia and adjacent areas (MURRIS, 1980).

Up to the top break of the Late Triassic continental deposits, the Toarcian indicates a return to marine sedimentation, witness to the global scale transgression which westward submerged the Central Arabian platform. The highstand sedimentation rate is important with clay, limestone and dolomite, elsewhere with more or less fine- to medium grained silty - sandy input; sandstones at the lower part would be possibly Early Toarcian, and gypsum at the topmost part, Aalenian. On the west, large deltaic displays underline the continental - coastal boundary. Middle to Late Toarcian outcrops are dated by ammonites and other marine fauna.

In Arabian Gulf surroundings (Kuwait, Qatar, Bahrain, United Arab Emirates), Zagros basin in Iran, till northern Iraq, more or less age controlled (overall by microfauna) limestones and limy shales, sometimes organic-rich, silty - sandy or dolomitic, are considered as equivalent to the Saudi Arabia Marrat formation (ALSHARHAN & NAIRN, 1997). In Kuwait, the "Jurassic reservoir" is within the Marrat formation (YOUSIF & NOUMAN, 1997).

Shallow silty clayey carbonates are recognised as Toarcian in central and SE Oman Mountains, Interior Oman and in Musandam Peninsula (BÉCHENEC *et al.*, 1993; LE MÉTOUR *et al.*, 1995). They are the signature of a quite continuous sea level rise on a stable carbonate platform, which started during Middle Liassic and span until the Kimmeridgian. The Dhofar High is still emerged.

In Oman, the Toarcian overlies unconformably the older deposits while the Marrat Formation and its equivalents are quite conformable in Saudi Arabia. An area of subsidence settles in the central part of the plate and will progressively accumulate sediments thicker than on the north and south borders. The settlement of this depocentre was probably related to the conjunction of two

factors. Probably but not exclusively, the uplift of the north and south plate borders (Mardin and Dhofar highs) which would be related respectively to the nascent doming between India and Gondwana, and to the uprightness of the Pamphylian furrow in central Tethys. The eustatic invasion of the central part of the platform, a world-wide Toarcian transgression, has certainly a stronger effect.

During Toarcian, the Arabian plate is a stable continental margin eastward bounded by the Tethys Ocean. Few normal faulting features in onshore Kuwait and Saudi Arabia gulf area (CARMAN, 1996) may be related to a feeble intraplate stress related to the incipient rifting of India from Gondwana or with the east Tethyan ridge (LE NINDRE *et al.*, submitted). Following the more or less marked Lower Jurassic gap-unconformity which is the result of both non-deposition and intense erosion, facies belts trend grossly W-E across the Arabian plate. The interpretation of the palaeoenvironments distribution of MURRIS (1980) is still consistent (AL-HUSSEINI, 1997) with, from the emerged Arabian shield, an eastward succession of arid flood plain, clastic and mixed platform, shallow carbonate to evaporite platform, facing the Zagros basin and the isolated platforms prior to the shelf slope and the Tethyan active spreading ridge.

III.10.- Ethiopia - Somalia - Gulf of Aden - Yemen

The Toarcian transgression has first a limited extent in Northern Somalia (LUGER *et al.*, 1994a), trending approximately E-W along the present day Gulf of Aden coast; only the Early Toarcian is testified with ammonites. But, it is currently admitted that the region was totally drowned during the Middle - Late Toarcian (BOSELLINI, 1989). The sedimentary sequences yield a great variety of facies belonging to several palaeoenvironments which differ in the several marine overlapped areas.

In the Ahl Mado (LUGER *et al.*, 1994a), Berbera and Borama basins, continental - fluviatile formations grade upwards, first into fluviatile - deltaic tidal flat and beach, then in shallow-marine environments, but still with siliciclastic facies (silty marls, marls and marly limestones); the terrigenous input is assumed to come from the emerged Arabian craton. Shallow-water carbonates extend on the whole area until the Abbay River (BEAUCHAMP & LEMOIGNE, 1974), Ogaden and Mandera Lugh basins; high energy carbonate and basinal silty-clayey facies settle respectively around emerged areas (Bur Aqaba High) and in the deeper parts of the shelf (BOSELLINI, 1989). Interbeded restricted evaporitic facies of shallow water or coastal plain cannot be more precisely dated than Late Liassic, because biostratigraphic data are missing. The Blue Nile basin and the Tigrai Mekele area still proceed with continental - fluviatile type sediments.

In south Yemen (east of Aden and south of Sana'a), the Toarcian should be present in fluvio-lacustrine shales and sandstones (BEYDOUN, 1989, 1997; BEYDOUN *et al.*, 1996; AL-THOUR, 1997). Easternmost Socotra platform yields similar sandstones facies (BIRSE *et al.*, 1997), which appear to have been deposited in channels at the mouth of tidaly dominated estuarine or deltaic system (SAMUEL *et al.*, 1997).

Likewise in the Sinemurian, the several areas of sedimentation are in relation with the active NE-SW and NW-SE rift structures which play both in parallel and orthogonal orientations with the present-day coast-line (BOSELLINI, 1989).

The "African Horn" has marine connections with the Tethys realm through the so-called "Pakistan portal" seaway; no direct connection exists with the central Arabian platform as it was sometimes claimed (BEAUCHAMP & LEMOIGNE, 1974; BEAUCHAMP, 1978).

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9.- MIDDLE CALLOVIAN (157 - 155 Ma)

J. THIERRY[1]

I.- MAIN FEATURES

A Middle Callovian map was already built in the Tethys Programme (ENAY *et al.*, 1993); the Peri-Tethys map concerns the same biostratigraphic interval: the subboreal - submediterranean Coronatum Zone.

Even if some differences remain for the definition of the upper boundary of the Coronatum Zone, and subsequently for the Middle - Late Callovian boundary (THIERRY *et al.*, 1997a), this unit is very well characterised by ammonite associations all over the north and south Tethyan margins and Peri-Tethyan cratonic areas. In spite of an increasing provincialism since the Aalenian - Bajocian and the settlement of endemic faunas, the Callovian is a time of world-wide ammonite migrations which allow very good correlations. Reliable biostratigraphic units are defined and available for West and NW Europe areas (THIERRY *et al.*, 1997a), Central and East Europe areas, Russian platform, Precaspian and Trans-Caspian areas (MELEDINA, 1988), North African margin, Levantine areas and Arabian platform (ENAY *et al.*, 1987; ENAY & MANGOLD, 1994).

Faunal Middle Jurassic provincialism is rather strong in Middle Callovian. On the far north of Peri-Tethyan areas a Boreal realm extends on Russian platform, Greenland and northern North Sea while a Subboreal province covers all the rest of the North Tethyan margin, from Iberia to the Turan plate; typical boreal - subboreal taxa never cross southward the Tethys hiatus. A Submediterranean province extends both on North and South Tethyan margins from Iberia to the Turan plate and from Morocco to Levantine areas, partially overlapping the Subboreal province. A Tethyan realm settles on the nearest borders of the Tethys ocean, while some endemic faunas exist in Levantine and Arabian platforms and margins. On the contrary to the Boreal - Subboreal faunas, the Submediterranean and Tethyan faunas largely extend northward. In spite of a marked provincialism, the resulting faunal mixing allows precise and reliable biostratigraphic correlations, even with Boreal faunas and Levantine - Arabian faunas; the former migrate southward onto Greenland - North Sea and Russia, when the latter migrate westward onto Maghreb, Apulian - Sicilian areas, South Iberia and France. Very good correlations are possible on all NW Europe, as far as Moesia, Crimea, Caucasus, the Russian platform and Trans-Caspian areas.

Ammonite faunas are most of the time present and diversified in every marine deposits; accordingly, the Middle Callovian map precisely illustrates the facies and palaeogeography of the standard Coronatum Zone. The distinction of the lower boundary of the selected time slice in the sedimentary pile is more often easy because of a quite generally strong facies turnover, from limestones to marly layers. Moreover, the Early - Middle Callovian boundary is often underlined either by a more or less important sedimentary gap which covers the uppermost Early Callovian and earliest Middle Callovian, or by easily distinguishable facies, often condensed with ferrugineous oolites, especially in the distal platforms of North, West and Central Europe. Sometimes, the upper part of the deep-seated Jason - Anceps Zone (Jason - Tyranniformis subzone) and the lower part of the overlying Athleta Zone (Phaeinum - Rota subzone) may be included in the map data because they have the same silty-clayey carbonaceous facies, constrained by the time span of the world wide transgression, overall in distal platforms and basins.

In areas with scarce or without ammonite data, brachiopods, large benthic foraminiferas, calcareous algae and echinoid faunas, are used as marker-beds in fine grained to shaly-clayish coastal facies (GARCIA *et al.*, 1996; THIERRY *et al.*, 1997b) or in shallow marine platform carbonates (ALMERAS *et al.,* 1997; BASSOULLET, 1997a and b). Rather good biochronostratigraphic resolution is possible at the zonal or substage accuracy.

Calcareous nannofossils (DE KAENEL *et al.*, 1996; GARDIN, 1997), dinoflagellate cysts (RIDING & IOANNIDES, 1996; FAUCONNIER, 1997), ostracods (BODERGAT, 1997) and small benthic foraminiferas (RUGET & NICOLLIN, 1997) are diversified; their stratigraphic range can be correlated with the accuracy of the Coronatum Zone or the total Middle Callovian, mainly in marine areas with silty-clayey carbonaceous facies of the NW European platform. Significant biostratigraphic correlations are possible with radiolarians (DE WEVER *et al.*, 1985) overall in the deep marine sediments of the Tethyan realm (radiolarite-type or deep carbonates - hemipelagic oozes facies) and sparse localities in the Boreal - Subboreal realm (DE WEVER & VISHNEVSKAYA, 1997).

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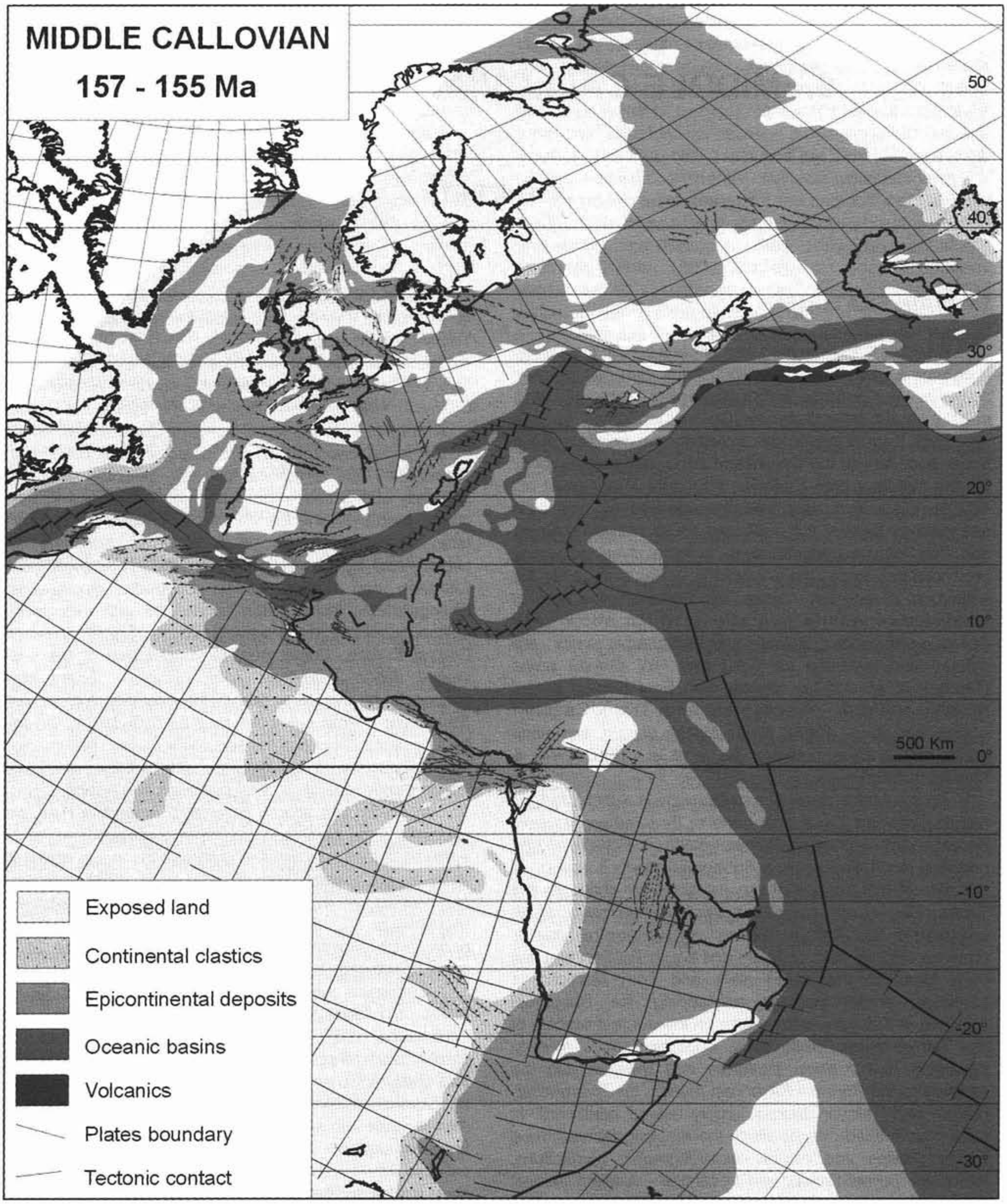

Fig. 9.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Middle Callovian.

The dating of continental or marginal shallow sediments (terrigenous or silty-clayey deposits of fluvio-lacustrine and fluctuating salinities environments) is based on spores and pollen associations. These are particularly useful in the intracontinental basins of the Saharan, African and Arabo-Nubian cratons. Associated with dinoflagellate cysts, they are the major dating elements of the coastal plain to shallow marine basins investigated by drillings all over the NW European platform.

The palinspastic reconstruction used is the Middle Callovian Tethys Programme map (Ricou, 1996), based on data acquired in the Pacific Ocean, at ODP site 801;

sediments with radiolarians of Early Callovian or latest Bathonian age overlie the upper oceanic crust and slightly predate the M38 anomaly (PRINGLE, 1992). Basic data for the Callovian - Oxfordian stages has been collected too on the Pacific crust where a complex signature of magnetic anomalies are assigned to M26 - M41 magnetochrons (SAGER *et al.*, 1998). However, the Callovian has not yet yielded a verified and commonly accepted magnetic scale (GALBRUN, 1995).

A succession of normal and reverse polarity, not yet completely accepted and calibrated by biostratigraphy or radiometry, is reported on several time scales (GRADSTEIN *et al.*, 1994, 1995; OGG, 1995; HARDENBOL *et al.*, 1998). This succession has been partly correlated to ammonite zones in South Poland (OGG *et al.*, 1991; OGG & GUTOWSKY, 1996), Spain (JUAREZ *et al.*, 1994, 1995) and Great Britain (OGG & COE, 1997). Subsequently, the base of the Callovian appears to correspond to the polarity subchron M36A. In the GRADSTEIN's time scale (*in* HARDENBOL *et al.*, 1998), the closer corresponding oceanic M-sequence for the Middle Callovian Coronatum Zone, would be the M38 - M39 magnetochrons.

There is no available radiometric data correlated with a European Callovian ammonite zonal scheme. In California, intra-Callovian volcanics, dated by ammonites, quote an age of 159 Ma (BOLES & LANDIS, 1984); ages of 161 Ma are given for the Bathonian/Callovian boundary and 160.5 Ma ± 0.2 for late Early Callovian beds with ammonites from Argentina (ODIN *et al.*, 1992). But, the correlations between South America and West Europe ammonite scheme are still uncertain.

A classical age of 166.8 ± 4.5 Ma is assigned to the sediments with radiolarians of latest Bathonian - Early Callovian age which overlie the Pacific oceanic crust at ODP site 801 (PRINGLE, 1992). An other set of data directly constrained by ammonite biostratigraphy comes from the Early - Middle Oxfordian of Switzerland (FISHER & GYGI, 1989): the Early Oxfordian Cordatum Zone may be near 149.2 ± 1.7 Ma; the Middle Oxfordian Densiplicatum Zone and the boundary between the Antecedens and Parandieri Zones are respectively 148.5 ± 1.6 Ma and 145.9 ± 1.8 Ma. Using the hypothesis of an equal duration for each ammonite zone within each stage, these data fit well with the ODIN's time scale which places the Callovian/Oxfordian boundary at 154 ± 5 Ma. Subsequently, the Middle Callovian Coronatum Zone is assumed to cover the 158-156 Ma interval. Such dating does not fit with the GRADSTEIN's time scale where the equivalent interval is much more older, between 161.3-160.8 ± 3.6 Ma.

Finally, referring to the ODIN's time scale, the Middle Callovian Peri-Tethys map is supposed to illustrate the palaeogeography between 157-155 Ma.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Laurasia and West Gondwana plates keep on their moving with a respectively southwards and northwards shifting of their eastern and western parts:

1.- the Trans-Caspian - Caucasus - Crimea - Dobrogea alignment continues to be controlled by the subduction of the Tethys oceanic crust which underlines the south border of the Iranian spur, the Sanandaj - Pontides island arc and the Moesian block. But, it is undergoing a compression phase which will amplify until the Oxfordian;

2.- the Pangaea is in an intense brake-up phase since the Late Bajocian. In the Central Atlantic, assumed to be at the moment of acquiring a mid-oceanic ridge lined by a narrow crust, the expansion is westward distributed along a rift zone between North America - Newfoundland margin and the Atlantic façade of Morocco; far to the west, it is time of the opening of the Gulf of Mexico. Active extension still plays between West European and Saharan cratons, widening the several furrows and basins not yet with oceanic crust along the Maghrebian transfer zone which makes the link between the Atlantic and Ligurian ridges (FROITZHEIM & MANATSCHAL, 1996; STAMPFLI & MARCHANT, 1997; ZIEGLER *et al.*, 2000). The coaval opening of the Magura basin, the Ligurian rift and the oceanic furrow off Corsica - Sardinia, all assumed to have oceanic crust which prolongs onto the Apulian block, contrasts with the compression which is henceforth undergone by the ancient Vardar Ocean, between Moesia and Ticza complex blocks. Passive thermal subsidence still plays in the Polish Trough and several NW Europe basins;

3.- the westwards propagation of Tethyan branches is active only in the Pindos - Olonos deep furrow, henceforth with continuous oceanic crust. The Bükk area is included in the complex Ticza block. The long deep Pamphylian basin still separates the Arabian promontory from the SW-Central Tethyan platforms;

4.- the border of the Arabian craton is still a passive margin with a wide and complex shelf, facing the Tethys ocean widely open to the east. The rifting between Arabia-Somalia and Indian plates is very active (KAMPUNZU & POPOFF, 1991; PLUMMER, 1996; VISSER & PRAEKELT, 1998; ZIEGLER *et al.*, 2000) driving to a major transgression and the widening of the shelf which connects with the Indian - Malagasy basins, through the Somalian and Ethiopian platform.

The main blocks which can be accounted for in the Middle Callovian are inherited from the Aalenian - Bajocian - Bathonian palaeogeography:

1.- The East European and the Russian platforms are the most important parts of the Laurasia; they are invaded since the Bajocian by a marine transgression which extends on low relief areas between the Fenno-Scandian shield and the Uralian highs; the Ukrainian and Stavropol shields are still emerged;

2.- the Precaspian basin and the Turan plate are drowned too, allowing marine connections with the Russian platform which is southward opened on the Great Caucasus marine basin. Mangyshlak and Kara Bogaz highs remain emerged;

3.- the Iranian block is separated from the Turan plate by the Great Caucasus basin;

4.- the wide Caucasus to South Crimea back-arc basin, still with abundant detritic input, is underlined by the Lesser Caucasus - Pontides island arc which volcanics and emerged ridges extend westward to the emerged Rhodope massif;

5.- the Moesian block, totally drowned except reduced high sectors, is still underlined by the North Tethys subduction zone;

6.- the succession of troughs and basins that border the Scythian platform and the Fenno-Scandian shield, from North Central Dobrogea to the Ergesund - Danish basin, through the Baltic platform, the Polish trough and the East Carpathian Gate, correspond to marine environments. The latter undergoes a passive thermal subsidence; it separates the huge north shields from the Bohemian massif which is still a large island. Possible temporary connections may exist with the shallow marine areas of the Russian platform;

7.- the Mid and SW European areas keep to enlarge their connections with the surroundings. As a consequence of permanent marine invasions on lowlands, they still more appear as a patch-work of shallow platforms, shoals, or more or less emerged islands separated by deeper platforms, basins and troughs. In the North Sea, the onset of the dome centre deflation, previously emerged during the Aalenian - Early Callovian interval, drives to marine incursions which follow a complex succession of differential subsidence on the margins and the centre of the dome;

8.- the north façade of the Iberian block shows unchanged connections with West Europe along the tectonic alignments of the future Bay of Biscay rift. Its Atlantic (Lusitanian basin) and south (Subbettic domain) rims still proceed with an extensive regime which is characteristic of these unstable Mediterranean areas. To the West, the Atlantic façade of Morocco registers too a transgression into the several basins which now face a symmetrical ocean with a mid-oceanic ridge (the "Atlantic Tethys");

9.- from Morocco to Tunisia, the Atlasic domain is no more below rifting conditions but only with active extension, stronger in the Tunisian part than in the Algerian and Moroccan parts;

10.- the Saharan areas and connected continental basins still register active extension and a coeval transgression;

11.- the present day shoreline of Libya and Egypt remains a passive margin; the distal part of the continental basins is in connection with the marine domain;

12.- the Levant and Arabia - Iran areas are still passive continental margins with strong asymmetric width of their shelf parts. As a consequence of the world-wide Callovian transgression, the latter is quite totally invaded, while embayments grow up eastwards on the former. A marine connection is proved between these shelves by mean of an E-W platform, north to the Arad high. On Arabia, the shore-line is westward moved, drastically reducing the emerged cratonic area which is underlined by huge deltaic complex;

13.- the African corner is similarly drowned by marine waters, coming both from Tethys and Malagasy domain, driving to a large shelf, which shoreline is far to the west onto the Anza basin - Blue Nile alignment.

II.2.- Palaeoposition of plates and blocks

The palaeoposition of plates is similar to Callovian Tethys map one (Ricou, 1996): there are no new available data nor computed new position of blocks and the palaeolatitude grid remains unchanged (Besse & Courtillot, 1991). The only modifications are the position of Iberia and Corsica - Sardinia blocks which are respectively west and NW shifted according to kinematics model, tectonic trends and facies distribution (Fourcade *et al.*, 1977; Canérot, 1991; Olivet, 1996; Vially & Trémolières, 1996; Verges & Garcia-Senz, 2000; Vera, 2000). The position of Moesia during Callovian is similar to the Tethys Programme map, in agreement with the palaeolatitudes data of Moesia (Surmont *et al.*, 1991; Tchoumatchenco *et al.*, 1992).

II.3.- Accuracy

The triple junction zone between Laurasia, Africa and the complex western central part of Tethys is considered as a "link zone" in the chain which allows to reconstruct the relative position of the several blocks within the sequence of palinspastic maps (Ricou, 1996). The classical Middle Callovian Atlantic fit is the oldest reconstruction, and the first in the Jurassic to be well enough constrained by a set of data (magnetic, biochronostratigraphic and radiometric) which ensures a good enough accuracy.

The biostratigraphic datings allow very good correlations within a high resolution frame based on several fossil groups.

Sequence stratigraphy provides very good correlations when the remarkable surfaces are very well constrained by reliable biostratigraphic data. The detailed succession and the hierarchy of the several order sequences and cycles for the Callovian are available overall in North Europe (Donovan *et al.*, 1993) and NW Europe (Rioult *et al.*, 1991; Graciansky *et al.*, 1993; Jacquin *et al.*, 1998; Stephen & Davies, 1998; Graciansky *et al.*, 1999; Guillocheau *et al.*, 1999; Robin *et al.*, 2000); they are compiled on the Jurassic Sequence Chronostratigraphy / Biochronostratigraphy Chart (Hardenbol *et al.*, 1998). Tentative correlations are possible with East Europe (Polish basin: Feldman-Olszewska, 1997a and b; Russian platform: Sahagian *et al.*, 1996) and the Arabian Gulf (Al-Husseini, 1997).

II.4.- General comments

The Middle Callovian map takes place at a major second order transgressive episode, initiated near the top of the Middle Bathonian; its peak transgression is at the Callovian/Oxfordian boundary and illustrates in Western Europe the first steps of the "North Sea cycle" (Jacquin & Graciansky, 1998). As a consequence, associated with a generalised rifting episode, epicratonic seas spread all over the Peri-Tethyan domains, drastically reducing the areas of shallow water carbonate platforms, which had a larger extent during the Aalenian, Bajocian and Bathonian. A

large part of the western European areas then shows a sedimentation dominated by shaly facies ("Marnes noires"). This important drowning is especially marked too on the African borders (Atlantic façade of Morocco, Saharan and Pelagian platforms, Egypt and Middle East, Arabian and Somalian - Ethiopian platforms) where carbonated sedimentation is very well developed on large and flat shelves, as well as on the complex system of platforms of the mediterranean "*Seuil lithosphérique*".

An other major palaeogeographic trend of the Middle Callovian is the deflation of the previous thermal doming in the North Sea (UNDERHILL & PARTINGTON, 1993). Initiated during the Late Toarcian - Early Aalenian, the maximum rise and emersion of the dome centre is reached in the Bathonian - Early Callovian. The Middle - Late Callovian corresponds to the beginning of deflation of the dome margins which leads to marine incursions in distal positions, while the centre continues to rise and produces paralic sequences on an irregular fault-controlled topography.

The Middle Callovian is time of major biogeographic events. From the latest Bathonian to the latest Callovian, the faunal exchanges are favoured and driven by a succession of transgressive episodes (third order transgressive regressive cycles; JACQUIN *et al.*, 1998) and distensive tectonic phases which enlarge or open seaways, both on wide and flat platforms or in depressed marine areas. For the first time since the beginning of the Jurassic, the Russian platform is totally drowned. Following the continuous transgressive trends, it is progressively invaded during the Bajocian and Bathonian, both from the North (Barents - Pechora basin) by Boreal and Subboreal faunas, and from the South (Precaspian basin) by Subboreal, Submediterranean and Tethyan faunas. The Greenland - Norvegian seaway ("North Atlantic Pass") opens again (DORÉ, 1991, 1992); it was formerly closed during the Bajocian - Bathonian by non-marine paralic sediments (major doming of the North Sea; UNDERHILL & PARTINGTON, 1993). The Atlantic and Central Tethys domains are connected through the Maghreb - Iberian transfer zone, the so-called "Hispanic seaway" ("Hispanic corridor"; ENAY, 1980; CARIOU *et al.*, 1985). Faunal migrations settle through the several newly drowned platforms on the African and Levantine borders onto the Arabian and Somalian areas; temporary connections may have existed too with the Caribbean areas and Western Tethys (CARIOU *et al.*, 1985).

The Middle Callovian is time when the *ammonitico rosso* facies is very well developed on the slopes, rises and shoals of many places of the Tethys domain. At the same time, radiolarite deposits reach a maximum extent, particularly in the deep sediments of the Tethys ocean and the so-called "Mediterranean *Seuil*" (FOURCADE *et al.*, 1996). They are too significantly present in the microfauna associations of medium to deep carbonates-hemipelagic oozes facies and shaly to clayey platform deposits, as far as on the Russian platform (DE WEVER *et al.*, 1985; DE WEVER & VISHNEVSKAYA, 1997). Such extent is reputed to be related to an abundant nutrient and an intense sea water circulation (oceanic currents and upwellings) associated with suitable conditions of silica preservation (DE WEVER *et al.*, 1996). It also coincides to a peak transgression and a carbon stable-isotope positive anomaly; the latter is interpreted as changes in global climate towards warmer and more humid periods, characterised by increased nutrient mobilisation and increased carbon burial (BARTOLINI *et al.*, 1996).

III.- DEFINITION OF DOMAINS

III.1.- Russian platform: Volga - Ural - Donetz - Ukraine

During the Early Dogger, major palaeogeographic changes occur on the Russian platform which is progressively drowned by world wide transgressions, driving to a large shallow-medium deep epicontinental sea. The first one, in Late Bajocian - Early Bathonian, flooded the majority of the previously emerged areas (VINOGRADOV, 1968): west Middle Volga, Volga - Kama, Ooka, Moscow and Upper Volga basins, onto the East Volga basin, the Pripyat, Dniepr - Donetz, Voronezh basins on the west and the Soukhona - Vichegda basin on the north. Henceforth, a wide shallow epicontinental sea extends on quite all the Russian platform, but no connection yet exists, neither with the boreal seas nor with the East European platform. The Fenno-Scandian, Ukrainian shields and Ural highs are still continuous emerged areas. Intermittent connections are assumed with the Scythian platform, Moesia and Dobrogea - Crimea - Pre-Caucasus areas; but, generally, the Ukrainian and Stavropol shields are still linked, building a huge emerged barrier. Opposite, large seaways exist on the Caucasus and Precaspian basins, and on the Turan plate.

The boreal seaway opens in Middle Callovian through the Sukhona - Vichegda, Mezen and Barents - Pechora basins; at the same time, the Russian platform epicontinental sea reaches its maximum extension between the huge Fenno-Scandian shield and the Ural highs - Kazakhstan plateau emerged areas. Direct connections between the Tethyan and Boreal realms are henceforth possible. A narrow and intermittent marine seaway (fine detrital coastal to shallow marine with fluctuating salinity deposits) opens between the Pripyat - Dniepr - Donetz basins and the Baltic platform - Polish trough. Intermittent connections may exist too with the Scythian platform, Moesia and Dobrogea - Crimea - Pre-Caucasus areas across the Ukrainian - Stavropol shields; large seaways continue through Caucasus, Precaspian basins and Turan plate.

In Middle Callovian, the sedimentation is detrital all over the Russian platform. Clays and shales are dominant; silts, sands and gravellites underline or surround the emerged areas. Scarce limy to marly intercalations, phosphatic concretions and ferruginous oolites are recorded in the Moscow and Sukhona - Vichegda basins. Fairly good biostratigraphic data are based on ammonites (OLFERIEV, 1986; MELEDINA, 1994), radiolarians (DE WEVER & VISHNEVSKAYA, 1997), nannofossils and small benthic foraminiferas. Sea surface palaeotemperatures of the Russian platform has been inferred from stable isotope measurements on belemnite rostra (RIBOULLEAU *et al.*, 1998); near 7°C in Late Callovian, they upgrade to 18°C in the Oxfordian. Such results, which run in the same direction as data collected in more septentrional areas

(King Karls Land, Svalbard; DITCHFIELD, 1997), change the traditional view of warm equable global climates in the Jurassic and suggest a latitudinal temperature gradient. The observed warming of the high latitudes would be due to the enlarged connections with the Tethys ocean, coeval with the palaeobiogeographic arguments based on the boreal and Tethyan ammonite realms.

The sedimentation is mainly eustatically controlled all over the Russian platform (SAHAGIAN *et al.*, 1996). But, in the Pripyat - Dniepr - Donetz basin (STOVBA *et al.*, 1996; VAN WEES *et al.*, 1996; STEPHENSON *et al.*, 2000), marly-sandy to sandy-limy shallow water Callovian sequences thin smoothly to the direction of the Donbass area, perhaps due to a reduced subsidence along the north border of the Ukrainian shield, but also because of later erosion (ULMISHEK *et al.*, 1994; STOVBA *et al.,* 1996).

III.2.- Turan plate

Wide transgressions occur in the Middle Jurassic. The Aalenian - Bajocian - Bathonian coarse detrital lacustrine, marshy or fluvial deposits, fine upwards to shales, with silts and clays with reduced but frequent coal seams. Clasts from sedimentary rocks are missing, suggesting strong erosion of an emerged uplifted near basement. The deposits are progressively marginally marine ingressed by fine shallow water to open shelf shales and marly limestones (North Ustiurt and South Mangyshlak depressions, Greater Balkan - Kopet Dagh and Karakum basins). Several areas remain undrowned in place (North Mangyshlak swell, Kara Bogaz high). The continental to coastal marine terrigenous facies henceforth extend on large areas of the Aral basin, along the emerged Kazakhstan plateau.

A first peak transgression occurs near the Late Bajocian - Early Bathonian. The whole East Caspian - Turan domain presents a submeridian alternation of shallow marine platform to coastal plain areas, separated by elevations.

After a short-term regression near the Late Bathonian - Early Callovian, a second peak transgression occurs in Early to Middle Callovian, marked too by a progressive shifting of the sedimentation from terrigenous to carbonates which sediment on platforms or open sea environments. The areas of shallow marine terrigenous carbonates extend on vast Turkmenian areas onto the North Caspian and Cis-Caucasus territories, connected to the south with the open marginal Tethys through the Lesser Caucasus. The areas of coastal terrigenous alluvial-delta sedimentation and the dissected denudation areas of the Turan plate decrease.

The stacking pattern, characterised by a low subsidence rate, is several time marked by unconformities, witness of the deepening and extensional movements which work since the beginning of the Jurassic (THOMAS *et al.*, 1999; LYBERIS *et al.*, 1998; LYBERIS & MANBY, 1999); they are known to cease during the Callovian, or Early to Middle Oxfordian. Thus, sedimentation is mainly driven by a gentle sea-level rise and a low subsidence regime. The stratigraphic succession is almost known from subsurface investigations (DAVOUDZADEH & SCHMIDT, 1981, 1983, 1984; VOLOZH *et al.*, 1997), but several areas like Mangyshlak (GAETANI *et al.*, 1998) give very good outcrops. The Aalenian - Bajocian age of the coastal plain and marshes sediments is based on palynology; the muddy to outer shelf sediments, sometimes with condensed levels and enriched in iron, are very well dated Callovian to Oxfordian by ammonites, brachiopods and foraminiferas.

III.3.- Scythian platform - Crimea - Black sea - Caucasus - and Precaspian areas

The Black Sea - Scythian platform - Caucasus belt has a complicated palaeotectonic and sedimentary Dogger history driven by the subduction-related magmatic belt of the Trans-Caucasus - Pontides back-arc activities (NIKISHIN *et al.*, 1998a and b, 2000). Thermal subsidence in the Fore-Caucasus basin (ERSHOV *et al.*, submitted), weak compressional deformations and extension in the Great Caucasus trough (KORONOVSKY *et al.*, 1987; PANOV & GUSHIN, 1987; OKAY *et al.*, 1994; ROBINSON *et al.*, 1996), orogeny in Crimea (VOZNESENSKY *et al.*, 1998), orogeny and uplifts in Pontides, take place during the Aalenian ("Mid-Cimmerian" orogeny). A major subductional magmatic belt is active in the Trans-Caucasus - Pontides area during the Bajocian. Intra-Bajocian to Early Bathonian inversion tectonics lead to orogeny along the Great Caucasus - Black Sea - Crimea belt (ROBINSON *et al.*, 1996; ROBINSON & KERUSOV, 1997). At the same time, the Scythian platform and several surrounding areas undergo a regional thermal subsidence in the Late Aalenian - Early Bajocian. While areas are uplifted and thrusted, several basins fill at different times with fine and medium clastic coastal to shallow marine deposits, coal-bearing in places: the Kalminksky basin in the Karpinsky swell area, the east Pre-Caucasus basin, the east and west Kuban basins in the west Pre-Caucasus area, and the South Crimea - Great Caucasus deep-water trough (NIKISHIN *et al.*, 1998a). The East European platform and the Russian platform emerged areas are the main sources of the clastic supply. Shallow-water carbonates and marls fill the Fore-Caucasus areas. The Middle Bathonian is reduced or badly documented, and the Late Bathonian - Early Callovian is generally a gap-interval witness of the culmination of the inversion tectonics.

A rift phase, associated with a world-wide transgression, begins with the late Early to Middle Callovian onwards, along the Great Caucasus - South Crimea trough, South Caspian areas as well as North Dobrogea; the southern margin of Moesia and the Pontides are concerned too. Although the several basins and uplifted areas are nearly identical, the sedimentation is shallow-water dominated. The Great Caucasus trough undergoes a rapid subsidence. At first, during the Middle Callovian - Early Oxfordian interval, silts, sands and sometimes conglomerates, characterise the rims; then it becomes clayey with carbonates on the slopes and in the furrows. In the Middle - Late Oxfordian, reef belts set around the deep parts and uncompensated sedimentation drives to the infilling of the basins (NIKISHIN *et al.*, 1998a and b, 2000). In the eastern Crimean Mounts terrane (VOZNESENSKY *et al.*, 1998), Middle Callovian deposits grade from nearshore to pelagic environments; a slope appears in Oxfordian, with reef massifs on the rims and debris flow deposits. The

beginning of collision between the Crimean Mounts and a salient of the Scythian platform is registered at that time. Opposite, the South Caspian basin and Kopet Dagh areas undergo acceleration of subsidence since the Callovian onwards; they are interpreted as an unified deep-water basin with thinned continental crust (NIKISHIN *et al.*, 2000; BRUNET *et al.*, submitted).

The Precaspian areas are the seaway which connect the Russian platform and the Caucasus trough. Shallow proximal to distal marine deposits fill the Central Precaspian depression; sedimentation is elsewhere shaly to sandy dominated, mainly fed by the emerged Stavropol shield.

III.4.- Teisseyre/Tornquist zone - Moesian platform

During the Middle Jurassic, tectonic events mainly related to the North Sea thermal doming, caused a complete reorganisation of the palaeogeography of the Polish basin and adjacent areas. Because of the regression of the sea from vast areas in North Germany, North Sea and Scandinavia which became emerged, the Polish basin was partly isolated from the rest of Europe since the Early Aalenian. Then, all along the Middle Jurassic a gradual sea level rise, coupled with a change in the basin geometry, resulted in the progressive expansion of the marine environments and the re-establishment of the communications with Western European marine areas and Tethys. Due to the widespread transgression, faunal exchanges between the Boreal, Sub-boreal Mediterranean and Tethyan realms, install all over the Polish basin domain, as far as on the south Baltic and the Russian platforms.

The epicontinental deposits of the Polish Middle Jurassic have been subdivided into transgressive-regressive cycles (FELDMAN-OLSZEWSKA, 1997b). These cycles illustrate the quite continuous sea level rise which, step by step, reaches its maximum of transgression near the latest Callovian - earliest Oxfordian boundary. The marine sediments, surrounded by fluviatile to coastal plain deposits, were first limited to the narrow gutter of the Mid Polish trough, when the Early Aalenian, then Middle Bajocian transgressions entered through the East Carpathian gate from the Tethys. Subsequently, the sea expanded to the SW then to the NE in Early Bathonian and Late Bathonian. Significant regressive episodes occurred in Early Bajocian, latest Bajocian, Middle Bathonian and Early Callovian. The last Middle Jurassic cycle almost brackets the Callovian, except its lowermost and uppermost parts. The transgression is maximum in the Middle/Late Callovian; sediments overlap the most of the Polish basin, extending onto the East European platform as far as the south Baltic areas (MAREK & GRIGELIS, 1998), while a communication opens to the east with the Dniepr - Donetz - Pripyat domains.

The sedimentation is distributed into two distinct sedimentological provinces where the biostratigraphic control is respectively given by megaspores and sporomorphs or ammonites, brachiopods, bivalves and agglutinated foraminiferas (DAYCZACK-CALIKOWSKA *et al.*, 1997; DADLEZ *et al.*, 1998). Coastal marine and shallow marine sandy-silty limestones, sometimes bordered by fluviatile deposits, underline the emerged Scandinavian shield; an hypothetical connection with the Danish basin may be traced. Shallow shelf carbonate-clastic, carbonate ramp and starved shelf environments, with varied facies of calcareous and dolomitic sandstones or siltstones, marls, calcareous mudstones and clays, extend far to the SW. Condensed deposits, the so-called "nodular bed" represent a starved basin deposit which extends quite everywhere at the moment of Late Callovian/Early Oxfordian peak transgression. Deep marly and clayey facies extend, from the North Germany basin to the East Carpathian, all along the Bohemian massif. The absence of coarse clastics and the direct present boundary between dry land and deep platform facies suggest either that the shoreline of this permanent neighbouring land mass was far away to the south, or that its relief was very low (MALKOVSKY, 1987; ZIMMER & WESSELY, 1996).

During the Middle Jurassic, the geometry, facies and palaeoenvironments deposits of the Polish basin, still indicate rifting conditions (KUTEK, 1994, 2000; STEPHENSON *et al.*, submitted) alike in the Liassic interval (N-NE - S-SW transtensive and E-SE - W-NW transpressive regime; sinistral component). The only differences are first a weakest subsidence rate, and second, that transverse faults delimit segments on the Polish graben, with specific palaeogeographic evolution (DADLEZ *et al.*, 1995; HAKENBERG & SWIDROWSKA, 1997; LAMARCHE *et al.*, 1998; LAMARCHE, 1999). The uniform subsidence pattern is still guided by the continuous activity of synsedimentary bounding faults, but the asymmetric-half graben basin is SW dipping in its northern part, and NE dipping in its southern part.

The Middle Jurassic and in particular the Callovian of the Ukrainian Carpathian foredeep is not very well documented; it should be included in the coastal plain sandy limestones and the shallow marine platform carbonates which upgrade into the Oxfordian on the southern rim of the Ukrainian shield (IZOTOVA & POPADYUK, 1996).

The SE prolongation of the Teisseyre - Tornquist zone is illustrated by the Dobrogea area where the Galati - San Georghe, Peceneaga - Kamena and Capidava - Ovidiu faults are still main tectonic features. The opening of the wide Tethyan - Ligurian Magura areas with a narrow oceanic ridge would be coeval with successive slips of the Dobrogea faults generating an eastward displacement of Moesia, overall along the Peceneaga - Kamena fault which is henceforth a key tectonic feature (HIPPOLYTE *et al.*, 1996; BANKS, 1997; BANKS & ROBINSON, 1997; TARI *et al.*, 1997). For that reason, these faults have been hypothetically prolonged to the NW, in continuation with the Polish trough faults, through the Magura areas. The differential uplift and subsidence of this area is reflected by the continuous propagation of these faults which separate the Predobrogean depression and the East European platform, from the North, Central and South Dobrogea (GRADINARU, 1984, 1988; BANKS, 1997; BANKS & ROBINSON, 1997; TARI *et al.*, 1997). Callovian deposits are known in the Predobrogean depression, and North and Central Dobrogea, showing respectively shallow marine carbonates and marly clayey facies (MOROZ *et al.*, 1997), carbonated to terrigenous shallow marine deposits and turbiditic-like to deep-water marls (GRADINARU, 1993); South Dobrogea is still an uplifted emerged area. Inversion

movements ceased in the North Dobrogea area at the end of the Middle Jurassic as shown by the resumption of carbonate sedimentation during the Late Jurassic.

As a result of the continuation of the differential subsidence of the Moesian platform, and in conjunction with a constant sea level rise, numerous areas are progressively drowned during the Middle Jurassic. At the end of this epoch, near the Middle - Late Callovian, the sea covers the majority of the complex Moesian area except several horsts of the Bulgarian side and the remnant central part of the Thracian - Rhodopes massif. The Early Jurassic intraplate rifting structures continue their existence during the Middle Jurassic but the general tendency is a limitation of the horsts areas and the enlargement of the grabens (TCHOUMATCHENCO *et al.*, 1989; TCHOUMATCHENCO & SAPUNOV, 1994). The principal fault sets (NW-SE, NE-SW, E-W), inherited from the Early Jurassic, persisted during the Middle Jurassic, but, a fundamental reconstruction in the main structures takes place during the Callovian: three blocks are differentiated. The west and the east Moesian carbonate platforms have a permanent tendency to positive movements; they are the site of shallow water carbonate sedimentation sometimes with coral build ups or restricted environments. The various facies are well dated by ammonites, brachiopods, smaller foraminiferas and spores and pollen. Between them, the Central Moesian basin is situated in a lower bathymetric position with a permanent tendency to subsidence; it is the site of pelagic sedimentation (SAPUNOV & TCHOUMATCHENCO, 1987, 1990; SAPUNOV *et al.*, 1985, 1988, 1991). Toward the end of the Late Callovian, the Nis - Trojan trough enlarged in the Eastern Balkanides, between the Moesian platform and the emerged central part of the Thracian - Rhodopes massif (TCHOUMATCHENCO *et al.*, 1989).

III.5.- Western Europe platform

The Middle Callovian palaeogeography of the west European areas is controlled by tectonic, eustatic and sedimentary events which are strongly linked. The map shows, respectively on the SW and on the SE, the first steps of the sea-floor spreading in the Central Atlantic (MASSON & MILES, 1984) and in the Ligurian Tethys (LEMOINE & GRACIANSKY, 1988), both initiated during the Late Bathonian - Early Callovian (ZIEGLER 1990; YILMAZ *et al.*, 1996). But, the two spreading systems are not yet connected; the area between Iberia and Africa is still occupied by a set of troughs, carbonated platforms and dry lands, shaped by a complex of E-W normal faults which delimitate tilted blocks, horsts and grabens (VERA, 2000). To the north, the onset of the North Sea dome centre deflation and the associated rifting (UNDERHILL & PARTINGTON, 1993), drives to progressively restore the marine communications between the Boreal and the Mediterranean realms, which were interrupted since the Aalenian - Bajocian mid-Cimmerian events. Hence, the whole Western Europe is below the threefold effects of the "North Sea Cycle", and the Central Atlantic and Ligurian Tethys spreadings (JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998); within such a first order eustatic cycle, the Middle Callovian map locates during one of the major sea-level rise of a second order transgressive - regressive cycle, which peak transgression is near the Callovian - Oxfordian boundary.

The sedimentation is dominated by medium clastic input in the north, overall sands and silts in coastal marine to shallow terrigenous platform environments, bordering the numerous emerged areas; finer sediments develop in distal environments over the majority of the central part of the considered area, whereas the carbonates are restricted to the south. The Middle Callovian palaeogeography strongly contrasts with that of the Aalenian - Early Bajocian and Middle - Late Bathonian - Early Callovian, dominated by shallow marine platform or ramp carbonate deposits, with frequent build ups, which corresponds to successive regressive episodes; the transgressive ones locate during the Late Bajocian - Early Bathonian and Middle - Late Callovian - Early Oxfordian.

On the Lusitanian - Atlantic margin of Iberia, the Middle Jurassic shales and carbonates, including Early Callovian, blanket the basin and show a relatively E-W simple deepening facies distribution consistent with a thermal relaxation (WILSON, 1988). The infilling of the basin spans all over the Aalenian - Callovian interval; its closure began in the Middle Callovian and was probably followed by emersions in the Late Callovian - Early Oxfordian (SOARES *et al.*, 1993). At the same time, in NE Spain, the Iberian platform undergoes a carbonated sedimentation which facies evolution indicates an increase of energy within shallowing upwards sequences; these are controlled both by eustatic variations and local influence of extensional synsedimentary faulting, either related to a post-rift passive thermal subsidence (FERNANDEZ-LÓPEZ *et al.*, 1996; AURELL *et al.*, 2000; SALAS *et al.*, 2000) or episodic rifting reactivation near the end of the Dogger (CANÉROT, 1989, 1991). Partial or total hiatuses and the development of condensed and more or less ferruginous layers underline a second order peak transgression covering the Late Callovian - Early Oxfordian interval. The Iberian massif is a permanent emerged but flat dry land with scarce marginal deposits except in some places of Asturias and Portugal.

The Middle Jurassic palaeogeographic evolution of the Pyrenees (CANÉROT, 1991; VERGÉS & GARCÍA-SENZ, 2000) and the Aquitaine basin (LE VOT *et al.*, 1996) is characterized by a low WNW-ESE rate of extension related to the early opening of the Atlantic Ocean. The Jurassic series were probably deposited during a phase of regional thermal subsidence in relatively stable conditions (BRUNET, 1984). In the Middle Callovian, the deposits, which make the transition between the Atlantic domain and the western border of the Dauphiné - Helvetic basin, show a deepening ramp succession in calm waters, grading from restricted-lagoonal and carbonated palaeoenvironments on the east, to shallow marine carbonate deposits, then deeper shelf marls and clays to the west (JAMES, 1998).

The Massif Central platform extends eastward until the Corbières - Provence and Corsica-Sardinia platforms, southward to the Dauphiné - Helvetic basin. In Ardèche and Cévennes, the border of the platform corresponds to a

set of tilted blocks, limited by NNE-SSW normal faults trending which extends until the Jura domain (ELMI, 1990). The proximal parts of the basins fill with rhythmical carbonated and marly successions (Middle Aalenian - Early Bajocian) that become more argillaceous in Late Bajocian; the so-called "Terres noires" facies started in Late Bathonian and extend until the Early – Middle Oxfordian. The sedimentation is elsewhere guided both by the sea-level variations (accommodation space) and the structural evolution (crustal extension and relatively rapid subsidence) of the basin and its surrounding carbonate platforms (GRACIANSKY *et al.*, 1993, 1998a, 1999; JACQUIN *et al.*, 1998); more or less condensed deposits with gaps, underline the Dogger - Malm boundary.

The Paris basin, the Schwabian platform and the Franconian basin, bounded by the Armorican, London-Brabant and Renish massifs, are permanent features. The existence of clastic deposits on the Armorican and London - Brabant borders may be indicative of the closeness of the shore line. First mainly occupied by carbonated platforms (Aalenian - Early Bajocian, Middle Bathonian, latest Bathonian - Early Callovian), the marine areas are several times drowned (Late Bajocian - Early Bathonian, Middle Callovian); they become deeper and more and more connected, showing a generalised limy-argillaceous facies (DUGUÉ, 1989, 1990, 1991; PELLENARD *et al.*, 1999). The carbonated platforms are restricted to the central Paris basin and Jura shelf, with a sedimentation characterized by numerous partial or total hiatuses, and by the development of condensed ferruginous oolite layers on its border during the Middle and Late Callovian, onto the Middle Oxfordian. The tectonic control of the sedimentation is still demonstrated but it decreases since the Aalenian onto the Kimmeridgian, coeval with an increasing and homogeneous subsidence speed and accommodation space (GUILLOCHEAU *et al.*, 1999; ROBIN *et al.*, 2000). Within the intracratonic basin, the Callovian is time of the turnover between the rifting episodes and the passive margin thermal flexure regime of the Ligurian Tethys. The more pronounced subsiding areas are near the major NW-SE fault zones but the previously NE-SW trends remain active in the centre of the basin. Such a fundamental reorganisation of the subsidence areas is the consequence of the major intra-plate "mid-Cimmerian" deformations which affect the whole Western Europe, since the Aalenian - Bajocian North Sea thermal event.

The Bohemian massif remains too a permanent emerged block south of the Hanover - Lower Saxony and Eastern Germany basins which undergo a marly and clayey sedimentation (KÖLBEL, 1968; WEITSCHAT & HOFFMANN, 1984). But, it yields erosional remnants of Late Callovian to Kimmeridgian deposits (MALKOVSKY, 1987) which suggest the existence of the so-called "Saxonian trough", located on NW-SE trending wrench faults, linking the NW Europe and the Tethys. Thus, Callovian transgressions may have entered the Bohemian massif, favoured by the world wide sea level rise and the West European extensional events.

The palaeogeographic evolution of the United Kingdom, North Sea and surrounding areas is totally controlled by the thermal events, the associated sea-level variations and rifting episodes of the "North Sea cycle" (STEEL, 1993; PARTINGTON *et al.*, 1993; JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998). Following the classic scenario (UNDERHILL & PARTINGTON, 1993), the initial rise of the dome centre starts in the Late Toarcian - Early Aalenian with evidence of shallowing ("Mid-Cimmerian unconformity"). The dome centre emerges during the whole Aalenian and Bajocian, driving to paralic hinterland and providing to the north the clastics of the Brent province; the deflation of the dome margin begins at the same time. It is generally claimed that from Early Bathonian to Early Callovian, the erosion rate of the dome centre remains high as it continues to rise with the development of paralic facies, fault control topography and volcanism, while the progressive subsidence of the dome margins leads to marine distal floodings. The volcanism regime may have span over younger times than Bathonian - Callovian (SMITH & RITCHIE, 1993); bentonite layers, certainly connected with the North Sea events, are precisely dated as Early Oxfordian within the clayey facies of the NE Paris basin and the "Terres noires" of SE basin in France (PELLENARD *et al.*, 1999). Whatever it be, the onset of the dome centre deflation takes place during the middle Late Callovian - Early Oxfordian, coeval with active volcanism, faulting of the heart of the dome, marine incursions along axial depressions of differential subsidence and on dome margins, lower erosion rate and less clastic supply. The Middle Callovian map locates at time of the onset of the first major extension in the northern North Sea (RATTEY & HAYWARD, 1993); the intense rifting phase of the whole area commenced in the Early Oxfordian and continued until the latest Oxfordian, onto its culmination in the Early-Mid Kimmeridgian (CLARCK *et al.*, 1993; LEPERCQ & GAULIER, 1996; THOMAS & COWARD, 1996)

The lithostratigraphic data and subsequent palaeoenvironmental interpretations compiled on the Middle Callovian map are based on numerous detailed or synthetic publications which cannot be all listed here and on several still published palaeogeographic maps (BRADSHAW & CRIPPS, 1992; ENAY *et al.*, 1993). The inland outcrops are very well dated by ammonites: Dorset (CALLOMON & COPE, 1995), Yorkshire - Cleveland basin (RAWSON & WRIGHT, 1995), Inner Hebrides - NW Scotland (MORTON, 1987, 1989, 1990, 1992a and b, 1993; MORTON & HUDSON, 1995). The offshore record is more sporadic, due to more or less extended hiatuses affecting the Middle - Late Callovian - Early Oxfordian interval and the scarceness of reliable biostratigraphic data, overall based on palynology and micropalaeontology. But, the documentation is very abundant because the North Sea areas has been widely investigated for petroleum research (BROWN, 1990) and taken as a model for sequence stratigraphy applications: Channel basin (HAMBLIN *et al.*, 1992); Hebrides and western Shetland shelves (STOKER *et al.*, 1993); Cardigan bay and Bristol channel (TAPPIN *et al.*, 1994); Moray Firth (ANDREWS *et al.*, 1990; STEPHEN *et al.*, 1993; CASEY *et al.*, 1993; STEPHEN & DAVIES, 1998); south, central and north North Sea grabens (HERNGREEN & WONG, 1989; VAN ADRICHEM BOOGAERT & KOUWE, 1997; DONOVAN *et al.*, 1993; PRICE *et al.*, 1993; GATLIFF *et al.*, 1994; KOCKEL, 1995); Viking

graben (MILTON, 1993; STEEL, 1993) and Ergesund - Danish basin (JOHANNESSEN & ANDSBJERG, 1993).

In spite of a generalised relative sea-level rise, at the mapped Middle Callovian time, the coastal-shallow marine terrigenous deposits of the northern North Sea and the distal shelves of the southern North Sea and inland United Kingdom are still separated; the Scottish massif and the Ring Köbing Fyn high formed an E-W elongated emerged area, crossed by the Central graben which is still occupied by continental and fluviatile deposits. The Shetland platform and several highs are connected too by non-marine facies. As a result, the only N-S direct marine connections would be through the Faeroe trough and Viking graben: on the one hand in a SE direction, through the Ergesund Danish basin and Scania, onto the Polish basin; on the other hand in a SW direction, through the Minches basin, Irish and Celtic Sea basins, onto the London Channel, Western Approaches and Yorkshire-Cleveland basins. In the latters, the shaly-clayey sedimentation partly corresponds to the so-called "Oxford Clays" or equivalent, which can be organic rich in some places (HUDSON & MARTILL, 1994; KENIG *et al.*, 1994; MARTILL *et al.*, 1994); the argillaceous input is reputed to be provided both by the numerous local emerged areas and the nascent Atlantic domain (DUGUÉ, 1989, 1990, 1991; PELLENARD *et al.*, 1999)..

III.6.- Maghreb (Morocco - Algeria - Tunisia) - Saharan areas

During the Aalenian - Early Bajocian time, the "mosaic episode", marked by general tectonic and sedimentary instability, is general over all North Africa (ELMI, 1996; KAMOUN *et al.*, 1999). The palaeogeography is elsewhere characterised by a strong contrast between deep basins and troughs (High and Middle Atlas, Tlemcen domain) which fill with marly sediments, often interrupted by gravity-flows: calcareous or even sandy turbidites, breccias. Carbonate platforms and sabkhas develop in North Sahara, Moroccan Meseta, Oran High Plains, Constantine block and the south of Tunisia; neritic and often ferruginous thin and condensed deposits set on the more prominent shoals. The transition to the basins is made by slopes characterised by marly limestones alternations. Alike in Liassic, the sedimentation is still highly controlled by active extension and differential subsidence; such a late stage of rifting is characterised by normal faults and tilted-block regime with a sinistral transcurrent component along a NE-SW trending in Atlasic domain (BEAUCHAMP *et al.*, 1996; PIQUÉ *et al.*, 1998; CHOTIN *et al.*, 2000; PIQUÉ *et al.*, 2000). But, compared to the Liassic, the subsidence rate is weaker (VIALLY *et al.*, 1994; BRACENE *et al.*, submitted; ELLOUZ *et al.*, submitted) and local movements due to gravity are supposed to play. A regional N-S extension induced by thermal subsidence establishes a horsts and grabens framework (Tataouine and Chotts basins, Medenine high) in Tunisia (BARRIER *et al.*, 1993; BOUAZIZ, 1995; BOUAZIZ *et al.*, 1996a and b, 1998, 1999; HLAIEM, 1998; HLAIEM *et al.*, 1997; PATRIAT *et al.*, submitted).

A "final homogenisation" stage is diachronous in the several Maghreb areas, ranging from Early Bajocian to Callovo-Oxfordian (ELMI, 1996). The basins became progressively less divided and they undergo a more uniform marly sedimentation which generally precedes the edification of extended carbonate platforms and shallow environments.

In the Atlas domain, the peak deepening is registered by thick marls and turbiditic deposits near the end of Early Bajocian or beginning of Late Bajocian (ELMI, 1996; ELMI *et al.*, 1998). The final homogenisation is attained in Early to Middle Bathonian in the Western Atlas where carbonate platforms develop with build-ups. Deep basins deposits disappear in the southern and western parts of the NW Maghreb while they continue in Algeria and Tunisia until Callovian, sometimes reduced and condensed (PEYBERNÈS, 1992; PEYBERNÈS *et al.*, 1996). Terrigenous input begins to spread northward, supplied by the NW Sahara. In Callovian, clays, silts and sandstones of prodeltas are widespread (BENEST *et al.*, 1991, 1995, 1997); they southward grade into continental and sabkhas deposits, north-bordered by outer platforms. A significant transgression invades the Lower Sahara basin, until the limits of the Oued Mya, Dahar, Tataouine, North Sahara and Ghadames basins; coastal plain to shallow terrigenous and carbonated platforms, sometimes with evaporites, extend far to the south (LEFRANC & GUIRAUD, 1990; BELHAJ, 1996; BOUAZIZ *et al.*, 1996a and b), while the southern parts of these areas remain continental, as well as the Murzuq basin. The fluviatile deposits of the Dahar basin yield an abundant and diversified plants imprints, witness of successive and rapid development of soil (BARALE *et al.*, 1998). Data on lithology in the Algerian Sahara (argillaceous - sandy series with ferralitic alteration products) and fossil record (rich filicean wood and vertebrate rests), denote a more humid climate than during the Liassic - Early Dogger times (BUSSON & CORNÉE, 1991).

In the Tlemcen area, the marly deposits persist during the Late Bajocian; shoals and horsts remained often prominent until they are drowned in the Early Bathonian (ELMI, 1996; ELMI *et al.*, 1998). In Callovian, the eastern part of the Atlas (Ouled Nails and Aurès) until Tunisia (North-South Axis and Tunisian Dorsale; PEYBERNÈS *et al.*, 1990; SOUSSI *et al.*, 1991a and b, 1999) and offshore Malta (BISHOP & DEBONO, 1996; ARGNANI & TORELLI, 2000), are occupied by pelagic basins (with marls, siliceous limestones and *ammonitico rosso* facies) north-eastward opened on the Tethys (KAMOUN *et al.*, 1999; SOUSSI *et al.*, 2000). Active extension is constantly recorded elsewhere, and underlined by breccias and shallow-water resedimentation in deeper areas (Ksour Mountains during the Bajocian; Tunisian Dorsale during the Bathonian). The peak deepening occurs only during the Callovian - Oxfordian in the North Tlemcen and Tell - Rif domains; from the Late Bathonian to Oxfordian, deep-sea fans develop in the external Rif (FAVRE *et al.*, 1991), while at the same time deltaic sediments infill the Middle Atlas and progress in the External Rif. The fine terrigenous input is widespread; sandy turbiditic-like deposits extend in huge submarine fans on the NW Maghrebian slopes while siliceous limestones and Ammonotico Rosso facies develop in the Tunisian Dorsale.

The Essaouira - Agadir and Tarfaya asymmetric basins extend northeastward (as far as South El Jadida),

witness of successive sea-level rises and moderate to feeble thermal subsidence related to the Atlantic opening (MEDINA, 1994, 1995; LABBASSI *et al.*, 2000; CHOTIN *et al.*, 2000). NW-SE syn-rift extension guided by N-NE - S-SW and E-NE - W-SW normal faults widen the basins which register carbonated-dolomitic to shaly facies of marginal marine to platform environments dated by rare ammonites, brachiopods and large foraminifera (PEYBERNÈS *et al.*, 1987; BROUGHTON & TRÉPANIER, 1993; MEDINA, 1994; MORABET *et al.*, 1998). Deeper marly facies, breccias and fans (FAVRE & STAMPFLI, 1992) indicate slopes facing the Atlantic which mid-oceanic ridge henceforth exists. Reduced continental fluvio-deltaic deposits, capped or interbedded by intraplate basaltic lavas and doleritic sills, which develop in the Bajocian - Bathonian (AMRHAR *et al.*, 1997), are recorded east of Essaouira, but there is no connection with the High Atlas continental detritics. These tectonic and magmatic activities belong to the initial stage of the oceanic accretion in the Central Atlantic.

III.7.- Egypt - Sudan - Libya

In Cyrenaica, Western Desert, Nile Delta and Sinai the Dogger is characterised by a sea level rise which initiates in Bajocian; several time interrupted by minor regressive episodes in Early and Late Bathonian, its peak transgression falls near the Middle Callovian. The normal faulting and the shear system which played since the Late Triassic on the North Egypt margin, progresses westward, driving to the individualisation of more synde-positional tectonic elements that will be active from Bathonian to Kimmeridgian/?Tithonian: the North Sinai, Misawag, Quattara, Shushan and Matruh basins (KEELEY & WALLIS, 1991). It is noteworthy that two tectonic frames interfere. The first is reliable to the activity of wrench faults that fragment the African margin in a parallel direction of the so-called "Trans-African Lineament"; such a SW-NE trending corresponds to the orientation of the continental-fluviatile basins and highs of the so-called "stable shelf" (KEELEY & WALLIS, 1991; ANKETELL, 1996; EL-HAWAT, 1992; EL-HAWAT *et al.*, 1996; SMITH & KARKI, 1996; KEELEY & MASSOUD, 1998; AYYAD *et al.*, 1998). The second controls the W-E normal faulting that formed half grabens all along the African margin during the Jurassic rifting (MOUSTAFA & KHALIL, 1990; GUIRAUD & BELLION, 1996; MOUSTAFA *et al.*, 1998; GUIRAUD, 1998; GUIRAUD & BOSWORTH, 1999).

The detailed stratigraphy and age of the formations is still obscure because they are only known in subsurface. Moreover, reliable biostratigraphic data are rare and facies upgrade and interfinger each other following the sea-level variations and the superimposed local-regional tectonic events (ABDEL AAL, 1990; HANTAR, 1990; JENKINS, 1990; KERDANY & CHERIF, 1990). Middle to Late Jurassic continental sandy-shaly formations are registered in the southern areas; they range from subaerial environment of deposition to supratidal and lagoonal evaporitic. Marine carbonated to shaly facies (mostly Middle Jurassic) have deposited in shallow low energy environment. Shaly-sandy formations (Middle Jurassic - mostly Callovian) are shallow marine and becomes more continental in character when they southward and westward interfinger with the clastic deposits. Limy facies (Bathonian? - mostly Callovian - to ?Kimmeridgian/?Tithonian) corresponds to shallow marine and low energy conditions which laterally grades toward the south and west into the previous listed deposits. Carbonated (dolomitic to calcareous with few interbeds of sandstones and shales) facies (Middle to Late Jurassic) are of more open marine origin and moderate energy environments.

The transgression is characterised by a widespread shelf carbonate deposition always clastic rich in its proximal parts (GUIRAUD *et al.*, 2000). In North Sinai (Gebel Maghara), the middle part of the deposits carries a Middle Callovian ammonite fauna and globigerinas associations; the latters indicate an open shelf until the Oxfordian - Early Kimmeridgian (EL-HEINY & MORSI, 1992). The transgression reaches the Gulf of Suez (Galala plateau; DARWISH, 1992) and extends all over the Nile Delta and the north Western Desert in Early Callovian; large foraminiferas allow to fairly precise dating of the Bathonian - Oxfordian interval (ABDELMALIK, 1981). In Mid Callovian quite all the "unstable shelf" is drowned, probably favoured by a rifting phase which signature is normal faults marking the shelf-slope transition along the Sinai and Cyrenaica borders. For the first time in Jurassic, the present coast line of Cyrenaica is overlapped by marine facies. The transition between non-marine ("Nubian" facies) to marginal shallow marine dominantly detrital ("Marmarica" facies) and deeper marine ("Jabal Al-Akhdar" facies) deposits, is well documented in subsurface by dinocysts and spores and pollen associations (THUSU *et al.*, 1988). They may extend onto the north rim of Sirt Gulf but not yet in the Sirt basin itself.

In the continental domain, the fluviatile Dakhla, Al-Kufra and Erdis basins are linked, certainly as a conse-quence of the Late Dogger base-level rise. Moreover, witnessed of their thick clastic infilling, it is inferred that large rivers, flowing north-eastward along the Farafra - Bahariya high, provide a significant clastic input for the mixed deltaic and marine facies (KLITZSCH & WYCSISK, 1987; HERMINA, 1990; KLITZSCH, 1990; KLITZSCH & SQUYRES, 1990; KLITZSCH & SCHANDELMEIER, 1990).

III.8.- Levant (Israel - Lebanon - Syria - Jordan)

The interfingering of late Early to early Late Callovian formations drives to a composite Middle Callovian reconstruction in Israel. The interval is represented by well dated alternating marls, limestones and marls, and marls and shales in the Judean desert and North Negev (HIRSCH *et al.*, 1998), by limy marls in the present day coastal plain and massive limestones in the central Israel (Judean embayment, Hermon and Galilee). In several parts of the present day coastal plain, the Oxfordian lies directly on the karstified Early Callovian, witness of emersion and erosion during Middle and Late Callovian. Post-Jurassic erosion suggests the continuation of the above cited deposits in south Negev and Sinai, as well as the whole Callovian in the Gevar'am infra-Cretaceous eroded canyon.

In Lebanon, the upper part of thick and massive limestones is supposed to include the Callovian interval

and therefore partly correlated with the Middle to Late Jurassic platform carbonate of Galilee (WALLEY, 2000). They witness the persistence of middle to outer shelf conditions and transgressive trends possibly until Late Callovian - Early Oxfordian, then followed by a regressive phase until Middle - Late Oxfordian. Coral build ups settle here and there in the Mount Lebanon and north Anti-Lebanon area; marls and shales interfinger with limestones in south Anti-Lebanon.

In Syria, the Middle Callovian is characterised in rare places of the Coastal Chain by brachiopods and larger foraminifera and in the Hermon by ammonites, despite the Late Bathonian - Early Callovian and Late Callovian - Oxfordian are well known (MOUTY, 1997a and b; SAWAF *et al.*, 2000). Alike in the present day coastal plain of Israel, the Oxfordian sometimes lies directly on the karstified Early Callovian, witness of emersion and erosion during Middle and Late Callovian. The entire Jurassic, is unknown in Euphrates graben, Sinjar trough (ALSHARAN & NAIRN, 1997; JAMAL, 1998) because of Early Cretaceous regional uplift and erosion. Dolomitic to marly limestones in the Mesopotamian foldbelt are doubtfully considered as partly Callovian (ALSHARHAN & NAIRN, 1997; JAMAL, 1998).

Western Jordan is still an emerged area and North Jordan a coastline where continental to restricted and nearshore varied deposits onlap the northern flank of the Rutbah high (BANDEL, 1981; ALSHARHAN & NAIRN, 1997).

III.9.- Central Arabian platform and Gulf area - Iraqi platform - Oman - Zagros basin

The synthesis of the central Arabian platform, surroundings of the Arabian gulf onto Iraq and Zagros foldbelt provide excellent overviews for the biostratigraphic data and the correlation of the formations, the palaeoenvironments interpretation and the geodynamic reconstruction of the whole Arabian plate (LE NINDRE *et al.*, 1990b; ENAY *et al.*, 1993; GRABOWSKI & NORTON, 1995; ALSHARHAN & NAIRN, 1997; AL HUSSEINI, 1997; YOUSSIF & NOUMAN, 1997).

The Middle to Late Callovian formations which outcrop in central Saoudi Arabia are thick and shaly to calcareous formations; representing highstand deposits, they extend to the north and the south of the platform and indicate marine intrashelf conditions with reefal palaeoenvironments at the top. Their age is well constrained by ammonite fauna. Resting above the Bathonian, the deltaic facies of the Wadi Ad Dawasir undergo a clear retrogradation which indicates that they have not been transgressed during the Callovian. Thus, they are considered as diachronic and contemporaneous of the Callovian formations; subsequently, they are reported on the Middle Callovian map.

In the Arabian gulf surroundings, Iraq and Zagros foldbelt, several formations which are dated mainly by mean of microfossils, illustrate the Callovian map with shallow to deeper marine facies and palaeoenvironments: they are mainly limy-marly and hemipelagic in Kuwait, Qatar, United Arab Emirates, Iranian Gotnia basin and Mesopotamian foldbelt - foredeep, while shallow carbonates sediment in South Zagros (GRABOWSKI & NORTON, 1995; AL-HUSSEINI, 1997). In Kuwait, the various Bathonian to Oxfordian interbedded limestones and bitumen are one of the three major Jurassic petroleum reservoirs.

In Central Oman, Oman Mountains and Musandam Peninsula, Middle Callovian exists within thick calcareous facies (LE MÉTOUR *et al.*, 1995). The Dhofar high is still a large emerged area.

During the late Middle Jurassic, facies belts continue to extend in a W-E direction across the Arabian plate, from a drastically reduced and emerged Arabo-Nubian shield, through successively alluvial plain, lower coastal plain, shallow mixed and shallow carbonate shelves. Such palaeogeography is inherited from the Aalenian - Bajocian - Bathonian. Following the negative or low subsidense rates in Late Toarcian and Aalenian (uplifting of the Arabian platform borders, period of lacuna), tectonic inversion and dominantly tectonic subsidence generate ramp and onlap systems which illustrate a major drowning of the Arabian plate during Bajocian to early Middle Bathonian. The consequence is the setting of an internal carbonate platform over Central Arabia with predominant north and south terrigenous inputs. During late Middle Bathonian - Early Callovian interval, a sea-level fall and a decreases of tectonic effects in subsidence drive to a major unconformity and sedimentary gaps (non deposition and erosion in incised valleys; GRABOSKI & NORTON, 1995) coeval with the setting of the huge Wadi ad Dawasir palaeodelta fan.

Over all the Arabian plate, the highest sedimentation rates occur during Middle and Late Callovian, mainly controlled by eustatism and thus leaded by sediment supply and accommodation space (LE NINDRE *et al.*, submitted). However, even if the Middle to Late Jurassic is characterised by a tectonic quiescence, normal faults system in the western gulf area enhanced the effects of subsidence (CARMAN, 1996). Mainly orientated in a parallel direction to the passive Arabian plate margin, they controlled the resulting sedimentation with a regional intraplate stress, coeval with the sea level rise.

III.10.- Ethiopia - Somalia - Gulf of Aden - Yemen

During the Bathonian(?) - Early Callovian until the Kimmeridgian, NE Africa is drowned as a result of a major transgression; between the Berbera - Borama and the Ahl Mado basins, the northern Somalian relief (Erigavo uplift) remains emerged (BEAUCHAMP, 1978; LUGER *et al.*, 1994a). The marine connections with the Arabian platform through Yemen are henceforth wide. The Middle Callovian is not biostratigraphicaly recognised within the thick calcareous deposits which extend all over Ethiopia, Somalia and Erithrea. However, the Bathonian - Early Callovian, the Late Callovian and the Oxfordian are mainly characterised by benthic foraminifera, calcareous nannofossils and scarce ammonites, overall in the Ahl Mado basin (METTE, 1993), the Danakil Alps and the Abbay River basin (GETANEH, 1981; REALE & MONECHI, 1994; SAGRI *et al.*, 1998). The Middle Callovian should exist in the marine carbonaceous facies.

In Central Ethiopia, the transgression gives rise to the Antalo Limestone which type area is the Mekele outlier in

Tigrai (BOSELLINI *et al.*, 1995; MARTIRE *et al.*, 1998), but which extends in several basins: Abbay River-Blue Nile (RUSSO *et al.*, 1994), Danakil Alps (SAGRI *et al.*, 1998). It is a shallow marine succession of marls alternating with limestones and sandstones, which grain supported (feeble but constant terrigenous supply or bioclastics - oolites) and marly micritic limestones facies, range from inner high energy to outer calm water ramp of the coastal and shelf areas (FICCARELLI *et al.*, 1975; TURI *et al.*, 1980). It is correlated with several formations of the continental margin of Somalia (BOSELLINI, 1989; LUGER *et al.*, 1994a), Yemen (SIMMONS & AL-THOUR, 1994) and the Gulf of Aden (BEYDOUN, 1989; BEYDOUN *et al.*, 1996). In the Abbay River basin (GETANEH, 1981), the sedimentation shows supratidal, intertidal and lagoonal environments on a tidal flat with fluvial, lacustrine and subtidal deposits which underline the emerged Nubian craton.

The sequence stratigraphy approaches (BOSELLINI, 1989; SAGRI *et al.*, 1998; MARTIRE *et al.*, 1998) reveal numerous omission surfaces with burrows and rock grounds with bioencrustations and iron-oxide coatings, related to submarine sediment starvation or diagenesis due to subaerial exposure. Many others discontinuity surfaces within the same facies are interpreted as the result of higher frequency cyclicity. But, as there is no evidence for synsedimentary tectonics, the sea level changes may be found mainly in eustasy or regional intraplate stresses, in spite of the active rift phase between East Africa and India, which break-up takes place near the middle Late Callovian boundary (BOSELLINI, 1989).

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10.- EARLY KIMMERIDGIAN (146 - 144 Ma)

J. THIERRY[1]

I.- MAIN FEATURES

An Early Kimmeridgian map located at the Subboreal Baylei Zone and its Tethyan/ Submediterranean equivalent Platynota Zone, was already built in the Tethys programme (CECCA *et al.*, 1993); for the Peri-Tethys Programme, a decision was taken to change this interval, because it is very often difficult to identify and several causes drive to uncertain correlations: marked faunal provincialism, important gaps underlining the Oxfordian/Kimmeridgian boundary, unfavourable facies as "sequanian" type for index fauna preservation. Moreover, considering our state of knowledge, the exact equivalence between the Submediterranean Platynota Zone and the Subboreal Baylei Zone is still doubtful and the biostratigraphic definition of the lower boundary of the Kimmeridgian is not yet ratified: biostratigraphic arguments demonstrate that the Oxfordian/Kimmeridgian boundary defined in the Boreal Realm falls in the middle part of the "Upper Oxfordian" as defined in the Tethyan realm (ATROPS *et al.*, 1993, HANTZPERGUE *et al.*, 1997; MATYJA & WIERZBOWSKI, 1995; SCHWEIGERT & CALLOMON, 1997; WIERZBOWSKI, 1991).

According to the International Commission on Stratigraphy, the Kimmeridgian must be used in its reduced signification (*sensu gallico*) and the "Kimmeridgian (*sensu anglico*)" would not be so long used as a stage name. Accordingly, the considered interval for Kimmeridgian map of the Peri-Tethys Programme corresponds to the Cymodoce Zone, which falls in the upper part of the "Lower Kimmeridgian" (*sensu gallico*) or the lower part of the "Lower Kimmeridgian" (*sensu anglico*). The choice of the Cymodoce Zone is better for correlations, on the one hand with the Submediterranean - Tethyan Hypselocyclum/Strombecki or Divisum - Herbichi Zones, and on the other hand with the Borealis Zone ("Kitchini beds" *auct.*), the equivalent unit of the Russian Platform zonal scheme for the Boreal Realm (MEZEZHNIKOV, 1988a; PROSOROVSKAYA *et al.*, 1995; HANTZPERGUE *et al.*, 1998a).

The palinspastic reconstruction used is that of the previously computed Early Kimmeridgian Tethys map (RICOU, 1996), based on palaeomagnetic data which fit as close as possible to the selected chronostratigraphic interval: Early Kimmeridgian An BS/M25 - An M24/M21, which embraces a huge interval comprised between 153 - 148 Ma (GRADSTEIN *et al.*, 1994, 1995; GALBRUN, 1995), with a confidence interval of ± 3 Ma.

There is no available radiometric data correlated with a Kimmeridgian ammonite zonal scheme. The closest data is located at the Early/Middle Oxfordian boundary (FISHER & GYGI, 1989), which fits very well with the ODIN's time scale, but not with the GRADSTEIN's time scale.

For the Kimmeridgian, alternative solutions are provided by the palaeomagnetic calibration. On the one hand, the similarity between the polarity sequence of Kimmeridgian - Tithonian deposits of the Subbetic province and the M-sequence of marine magnetic anomalies, coupled with precise biostratigraphic control, allows assignment of the following ages: the Kimmeridgian/Oxfordian boundary falls either within or slightly after the M25 anomaly (OGG *et al.*, 1984; GRADSTEIN *et al.*, 1994, 1995), or near the end of the M25 anomaly (GALBRUN, 1995). The Early/Late Kimmeridgian boundary (top Cymodoce Zone) is near the end of M24 (OGG *et al.*, 1984), therefore, the Kimmeridgian/Tithonian boundary correlates with the end, or near the end of the M23. On the other hand, the Oxfordian/Kimmeridgian boundary in the Tethyan realm (base of the Platynota ammonite Zone), which was previously assigned to the top of M25n polarity chron (GRADSTEIN *et al.*, 1994), would be henceforth assigned to the older M25r chron (OGG & GUTOWSKY, 1996).

A set of noticeable discrepancies appears when comparing the biostratigraphic data and the subsequent uncertainties of the Oxfordian/Kimmeridgian boundary, the available magnetic polarity successions, the defined magnetochrons and the corresponding radiometric ages. Following both the GRADSTEIN's time scale (1994) and the GALBRUN's magnetic succession (1995), the Cymodoce Zone interval (which embraces the Hippolytense, Lothari, Tenuicostatum and Lothari subzones) may be between 154 - 152.5 Ma, with a bar error of 3.2 Ma.

On the ODIN's time scale (1994), the lower boundary of the Kimmeridgian is poorly documented and located at 146 Ma; the upper boundary, likewise poorly documented, is located at 141 Ma but would be 4 Ma older. Thus, referring the ODIN's time scale, if all ammonite zones or subzones have an equal duration, the concerned time slice for the Kimmeridgian Peri-Tethys map may be

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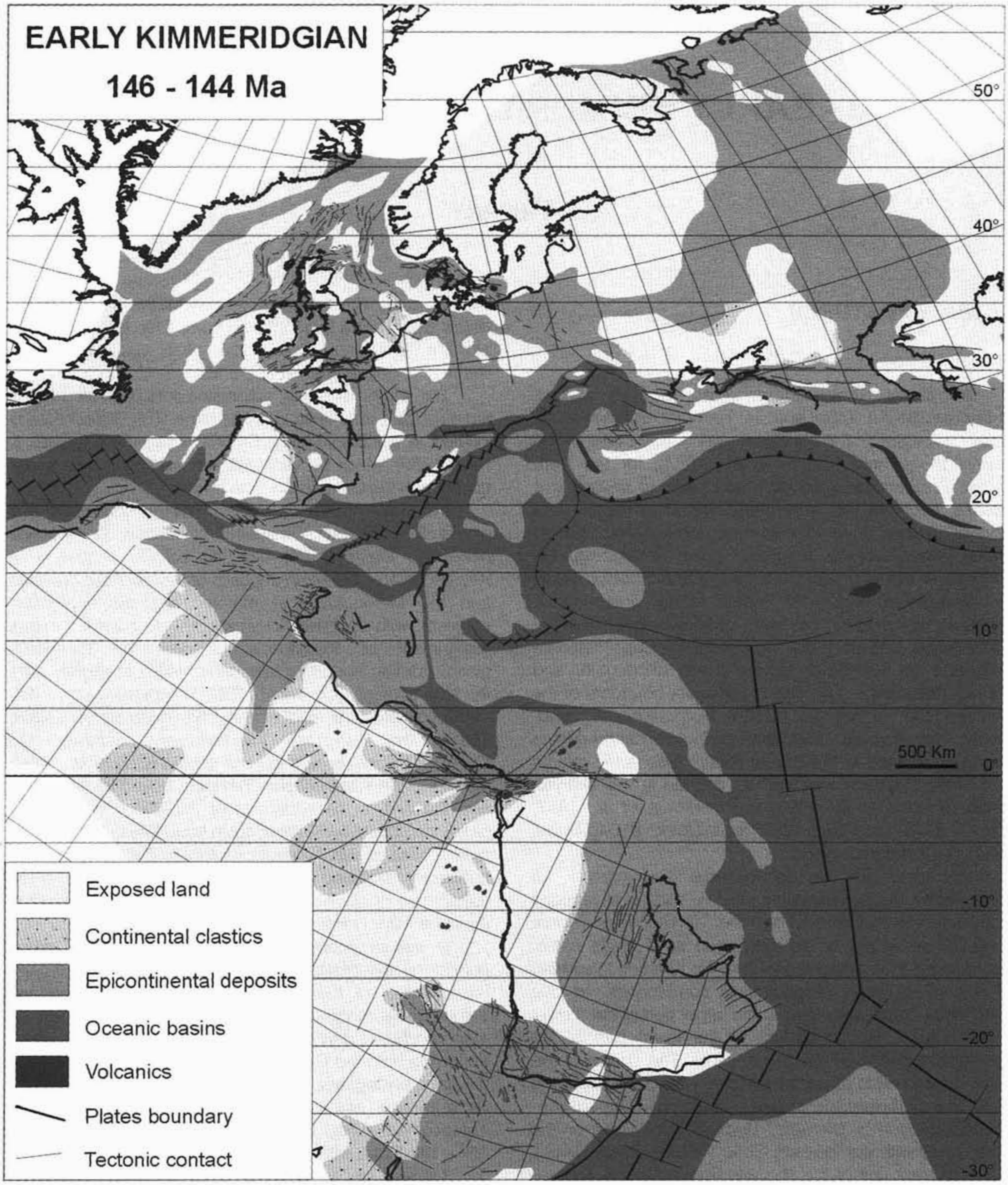

Fig. 10.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Kimmeridgian.

around 144-145 Ma. If not, for example considering the GRADSTEIN's scale where ammonite zone may have shorter or unequal duration, it would be between 152.5-154 Ma. When reliable biostratigraphic data are missing, such a difference would be kept in mind for ever estimation or precise location of events within the Kimmeridgian stage.

Finally, considering all the available data, but giving the preference to the ODIN's time scale, the Early Kimmeridgian Peri-Tethys map is assumed to be located between 146-144 Ma.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Following an unchanged anticlockwise rotation, Laurasia and West Gondwana plates keep on a respectively southward and northward shifting displacement of their eastern and western parts:

1.- the Transcaspian - Caucasus - Crimea - Dobrogea alignment continues to be controlled by the subduction of the Tethys oceanic crust which underlines the southern border of the Iranian block, the now reduced Sanandaj - Pontides island arc and the Moesian block. Since the Oxfordian, onto the Early Cretaceous, compression movements are registered in several places of the Scythian platform;

2.- the brake-up of the Pangaea is once more in an intense phase. The Central Atlantic is henceforth widely opened with an active mid-oceanic ridge in constant drifting activity, producing a NW-SE expansion between the North America - Newfoundland and Africa margins. Through the maghrebine transfer zone, the central Atlantic ridge is supposed to be joining the Ligurian ridge, off Corsica - Sardinia and the complex triple junction point in Apulia, between Africa, Laurasia and a set of unstable blocks. The unceasing and coeval opening of the Ligurian - Magura basins, and the oceanic ridge off Corsica - Sardinia, contrasts with the compression which has henceforth an effect on the ancient Vardar Ocean, between Moesia and Ticza complex blocks, first signs of the future latest Jurassic Dinaric - Hellenic obduction. In numerous NW European basins (Polish Trough, North Sea, Paris basin, Iberian platform), the rifting events progressively replace the diachronous and local passive thermal subsidence which played episodically since the Middle - Late Liassic and Dogger;

3.- there is no evidence that extension came to an end in the SW Central Tethys complex platforms. The westwards propagation of Tethyan branches is still active in the Pindos - Olonos furrow, henceforth with continuous oceanic crust. The long deep Pamphylian basin still separates the Arabian promontory from the SW Central Tethys platforms which are still linked to Apulia, Sicily - Malta escarpment and the Pelagian shelf;

4.- the border of the Arabian craton is still a passive margin with a very wide shelf east facing the Tethys. The oceanic ridge and the rifting between Arabia - Somalia and India has been very active since the Callovian; the plates are henceforth completely separated. A major transgression drives to the widening of the Somalian - Ethiopian shelves which are connected with the Indian - Malagasy basins.

The several blocks which are accounted for in the Early Kimmeridgian do not strongly differ from that of the Callovian palaeogeography:

1.- the East European and the Russian platforms are still the most important stable parts of the Laurasia; they continue to slowly subside and be covered by the Late Jurassic seas which maximum extension is reached in the Late Kimmeridgian. Numerous marine connections repeatedly exist throughout the Kimmeridgian the Earliest Tithonian, through a more or less E-W permanent seaway between the Fenno-Scandian shield and the Ukrainian high;

2.- marine connections still exist too between the Precaspian basin - Turan plate and the Russian platform which is southward opened on the Great Caucasus marine basin. Mangyshlak and Kara-Bogaz highs stay emerged and associated with the Kazakhstan plateau and the south Ural highs;

3.- the eastward ending of the Great Caucasus basin is unknown, but it probably still separates the Iranian block from the Turan plate. The Lesser Caucasus - Pontides island arc underlines the Great Caucasus - South Crimea back-arc basin which still undergoes active extension and fills with turbiditic facies; the low relief and coastal areas, the shallow platforms and troughs of the Scythian platform rim which have persisted since the Callovian - Oxfordian are finally quite all drowned;

4.- the Moesian block, including the Rhodope massif is still south underlined by the northern Tethys subduction zone; it is quite totally drowned too, except hypothetical high central parts which may prolong the western Pontides. The opening of the Carpathians - Magura basin, coupled with a strike-slip faults activity and the eastward rotation of Moesia, initiate compression in Crimea, associated with volcanism in Dobrogea;

5.- on the borders of the Ukrainian and Fenno-Scandian shields, the transgressions enlarge the marine areas of the troughs (Polish and east Carpathians) and basins (Baltic) which are occupied by shallow to deeper epicontinental seas, while the isolated emerged massifs (Bohemian) show a maximum reduction. In place, the sedimentation is enhanced by a reactivated rifting and active extension;

6.- the Mid and SW Europe similarly undergo constant marine invasions; it appears as an archipelago of reduced emerged islands separated by shallow platforms, shoals, deeper platforms, basins and troughs, shaped by reactivated faulting in a rifting context. Similarly, in the North Sea, the continued dome-wide deflation associated with rifting phases, drives to progressive flooding of axes of differential subsidence and adjacent margins and platforms;

7.- the northern façade of the Iberian block shows unchanged connections with West Europe along the tectonic alignments of the future Bay of Biscay rift. Its Lusitanian rim is still a passive margin. while its Mediterranean border is part of the Maghrebian transfer zone which henceforth joines the Atlantic ocean and the Ligurian ridge;

8.- the Atlasic domain, from Morocco to Tunisia, is still below rifting conditions with active extension, still stronger in the Tunisian part than in the Algerian and Moroccan parts;

9.- the Saharan areas and its marginal to continental basins continue to register active extension, coeval with a transgression which extends the

evaporitic and marine conditions to a maximum for the whole Jurassic;

10.- the present day shoreline of Libya and Egypt remains too a passive margin which undergoes a wide transgression; the distal parts of the continental basins are in connection with the marine domain, while their proximal parts develop evaporitic environments;

11.- the Levant and Arabia - Iran areas are still passive margins; the asymmetric width of their shelf parts reaches a maximum. The former is narrow and shaped by several embayment or shoals; in the future, it would become a continental edge. The latter is a wide ramp between the extremely reduced Arabo-Nubian shield - Arad high and the Tethys. On Arabia, since the Callovian - Oxfordian, the shore-line continued to move westwar, drastically reducing the emerged cratonic area which is still underlined by huge deltaic complex. Volcanic events take place in Levant platform, southern Egyptian and Libyan craton;

12.- the African corner is henceforth completely separated from the Indian plate; it is widely flooded too by marine waters, coming both from Tethys and Malagasy domain, driving to a large shelf, which shoreline is far to the west up to the Anza basin - Blue Nile alignment.

II.2.- Palaeoposition of plates and blocks

There are no new available data nor computed new position of blocks and the palaeolatitude grid remains unchanged (Besse & Courtillot, 1991). The palaeoposition of plates is taken from the Kimmeridgian Tethys map (Ricou, 1996) where North America and Africa are fitted along the M25 anomaly (Klitgord & Schouten, 1986), which constrains the width of the Atlantic ocean. Compared with the Tethys maps, a decision has been taken to add oceanic crust without sediments in the several furrows where drifting would exist (central Atlantic, Iberian - Maghrebian transfer zone, Corsica - Sardinia - Ligurian - Magura and Pindos - Olonos furrows. Alike for the previous Jurassic maps, the only outstanding modifications of several microplates are the position of Iberia and Corsica - Sardinia blocks which are respectively W and NW shifted according to kinematics model, tectonic trends and facies distribution (Fourcade *et al.*, 1977; Canérot, 1991; Olivet, 1996; Vially & Trémolières, 1996; Verges & Garcia-Senz, 2000; Vera, 2000). The position of Moesia is not modified following the Tethys Programme maps, in agreement with the sequence of palaeolatitudes data obtained in Romania and Bulgaria (Surmont *et al.*, 1991; Tchoumatchenco *et al.*, 1992).

II.3.- Accuracy

The classical Atlantic fit is the best available reconstruction and the only in the Jurassic to be well enough constrained by a set of magnetic, biochrono-stratigraphic and radiometric data. Thus, except the few modifications introduced and already discussed, the palinspastic reconstruction has a rather good accuracy. Some problems will always remain near the so-called "triple junction zone" between Laurasia, Africa and the complex western central part of Tethys. Such a "link zone" in the chain which allows to reconstruct the relative position of the several blocks within the sequence of Jurassic palinspastic maps (Ricou, 1996), would need new investigations. The selected palinspastic model strongly differs from that proposed for the Oxfordian (Stampfli *et al.*, 2000) where on the one hand the oceanic basins are much more developed than in the chosen reconstruction, and on the other hand the tectonic units which compose the W central Tethys blocks are differently developed and associated (Stampfli, 1993, 1996; Stampfli & Mosar, 1998; Stampfli *et al.*, 1998a and b, 2000).

Fairly good biostratigraphic datings within a high resolution frame are henceforth possible in spite of a strong provincialism (Hantzpergue, 1993). Very precise ammonite zonal schemes and correlations are established, on the one hand between Tethyan - Mediterranean and Boreal faunas, and on the other hand between West European (Euro-Boreal - Mediterranean) and Russian platform faunas (Hantzpergue *et al.*, 1997; Hantzpergue *et al.*, 1998a and b). Several microfossil groups are currently used in parallel scales, overall in NW Europe, including the North Sea domain and the Russian platform: radiolarians (Vishnevskaya, 1995; De Wever & Vishnevskaya, 1997; Vishnevskaya *et al.*, 1999), calcareous nannofossils (Riding & Ioannides, 1996; Gardin, 1997), dinoflagellates (De Kaenel *et al.*, 1996; Fauconnier, 1997) and small benthic foraminiferas (Ruget & Nicollin, 1997; Samson, 1997). Large foraminifers (Bassoullet, 1997) are frequently the only biostratigraphic data available in the widely developed carbonated platforms of the southern Tethyan and Peri-Tethyan areas.

Sequence stratigraphy provides too very good correlations when depositional sequences and transgressive/regressive cycles are constrained by reliable biostratigraphic data. The detailed succession and the hierarchy of the several order sequences and cycles for the Kimmeridgian are available overall in Northern Europe (Wignall & Hallam, 1991; Wignall, 1991; Donovan *et al.*, 1993; Partington *et al.*, 1993a and b; Price *et al.*, 1993; Rattey & Hayward, 1993; Steel, 1993) and NW Europe (Rioult *et al.*, 1991; Tyson, 1996; Jacquin & Graciansky, 1998; Jacquin *et al.*, 1998; Stephen & Davies, 1998; Guillocheau *et al.*, 2000; Robin *et al.*, 2000; Aurell *et al.*, 2000); they are henceforth compiled on the Jurassic Sequence Chronostratigraphy / Biochronostratigraphy Chart (Hardenbol *et al.*, 1998). Tentative correlations have been proposed with the Russian platform (Sahagian *et al.*, 1996) and the Arabian Gulf (Al-Husseini, 1997).

II.4.- General comments

As a consequence of successive opening phases in W central Tethys and Central Atlantic, the tectonic regimes governing the evolution of the Peri-Tethyan platforms changed repeatedly in conjunction with the permanent sinistral clockwise rotational - translation of Eurasia relative to Gondwana. The maximum intensity

of the tectonics will be progressively reached during the Late Jurassic onto the Early Cretaceous. The main resulting palaeogeographic evolution will be the development of new divergent plate boundaries, in the Atlantic between North America and Africa, and in the Indian Ocean between Africa - Arabia and India; a subsequent rifting regime is reactivated quite all over the cratonic Peri-Tethyan areas while the west central Tethys micro-plates undergo a progressive reorganization. In NW Europe (NE Iberian platform, Aquitaine and Paris basins, Western Approaches and Channel basins, North Sea and adjacent areas, Polish trough), North Africa (Atlantic border of Morocco, High Atlas and Maghrebian platform, Tunisian trough, Chott and Tataouine basins), Egypt and Levant margins, Arabian and Somalian - Ethiopian platforms, a Late Jurassic - Early Cretaceous stress activation is evident by the onset of rapid subsidence in rifting basins. In several places, everywhere in West Europe, the rifting events seam to be controlled by the reactivation of Permo-Carboniferous wrench faults.

Within a eustatic framework, the Early Kimmeridgian map locates near the outset of the major sea level rise of the whole Jurassic; it is near equivalent eustatic-palaeogeographic conditions as the Late Sinemurian, Early Toarcian and Middle Callovian ones, but it corresponds to a moment of not the same tectonic and probably climatic conditions. It illustrates outstanding features which are consequences of the worldwide transgression: the increase of faunal exchanges; the development of shallow carbonate platforms and carbonate deposition in the pelagic domains; the sedimentation of potential kerogenous source rocks in basin settings.

Kimmeridgian Boreal ammonites may migrate as south as the Aquitaine basin and Iberia while Mediterranean - Tethyan ammonites can be found as far north as the United Kingdom territories and Greenland (HANTZPERGUE, 1993); migrations are described too between W-NE-central Europe and the Russian platform (HANTZPERGUE *et al.*, 1998a and b). Such faunal exchanges introduce foreign taxa within the several regional biota and allow reliable correlations between the corresponding deposits and areas.

In the north Peri-Tethyan margin, carbonate platform environments dominate from central Iran to Iberia while terrigenous-marly-clayey facies are dominant to the north. The Mediterranean seuil, Apulian - Dinaric and Greece - Turkey territories are characterised by isolated carbonated platforms. The whole south Tethyan margin is underlined by more or less extended carbonated platforms which covered wide domains in Somalia - Ethiopia and Arabia, the former being more shaly than the latter. Such a development has begun in the Late Aalenian - Early Bajocian but, since that time, the carbonate factory had progressively changed; first dominated by temperate bioclatic crinoidal - molluscs limestones, they are progressively replaced by tropical chlorozoan - corals - sponges facies including huge build-ups in the Middle Bathonian - Early Callovian. Their maximum development is reached episodically in Middle - Late Oxfordian - Early Kimmeridgian and Tithonian. The carbonate production is sufficiently abundant to keep up and balance any increase in accommodation space, resulting of repeatedly subsidence and sea-level rises. The climatic evolution should have played too an important part on any increase in accommodation space, build-ups organisms and bioclasts producers, and subsequently on carbonate sedimentation.

Significant deposition of organic matter and changing balance between intracratonic Peri-Tethyan/isolated Tethyan platform carbonate deposition and the pelagic domain is too a main feature of the Kimmeridgian which is time of high atmospheric carbon dioxide level. The stratigraphical distribution of the organic rich facies is younger, strongly developed within the successions and span over much more time in the North Sea and West European areas (Late Kimmeridgian) than in the Russian platform where it is older (late Early Kimmeridgian) and reduced within the sedimentary pile (HANTZPERGUE *et al.*, 1998a and b). The important organic deposition in the Kimmeridgian seems to be balanced by increasing carbonate deposition. Since the Early Dogger, the oceanic domain became a more and more important site of accumulation of micritic pelagic limestones because of the moving of the carbonate-producing planktonic micro-organisms from the shallow epicontinental seas into the oceanic domain (DE WEVER *et al.*, 1996; BAUDIN & HERBIN, 1996). Such evolution with a culmination during the Kimmeridgian - Tithonian, is linked to palaeoecological, geochemical and physical modifycations of the sea water (BARTOLINI *et al.*, 1996; WEISSERT & MOHR, 1996). On the one hand, the nanno- and microfossils assemblages suggest a decreasing nutrient flux all along the Tethyan margins and Peri-Tethyan platforms. On the other hand, the carbon isotope record gradually decreases from the Oxfordian throughout the Kimmeridgian and the Tithonian; the Oxfordian fluctuations may be linked to periodic unbalance between the burial rates of carbonate and organic carbon production, while the more discrete record in the Kimmeridgian and Tithonian may be linked to a stabilization of the partitioning between the organic and mineral carbon.

III.- DEFINITION OF DOMAINS

III.1.- Russian platform: Volga - Ural - Donetz - Ukraine

A vast shallow N-S orientated epicontinental sea extends on all the Russian platform since the Callovian; its limits undergo feeble modifications during the Oxfordian (VINOGRADOV, 1968). At the onset of the Kimmeridgian, the boreal seaway is widely opened through the Sukhona - Vichegda, Mezen and Barents - Pechora basins, separating the Fenno-Scandian shield and the Ural highs - Kazakhstan plateau emerged areas. A narrow and shallow marine seaway is installed between the Pripyat - Dniepr - Donetz basins and the Baltic platform - Polish trough. Direct but reduced connections exist too between the Tethyan and Boreal

realms through Caucasus, Precaspian basins and Turan plate. The Ukrainian and Stavropol shields are linked, building an emerged barrier; as a consequence, no direct connections are possible with the Scythian platform, Moesia and Dobrogea - Crimea - Precaucasus areas.

Marine deposition occur in the majority of the sub-basins recognised over the Russian platform: the west Middle Volga, Volga - Kama, Ooka, Moscow and Upper Volga basins in the central part; the east Volga basin to the SE; the Pripyat, Dniepr - Donetz, Voronezh basins to the SW and the Soukhona - Vichegda, Mezen and Barents - Pechora basins to the NE. The revision of ammonite biostratigraphy of several Late Jurassic type sections in the Middle Volga basin (HANTZPERGUE *et al.*, 1998a and b) provides henceforth the best correlations between the zonal schemes of Europe (HANTZPERGUE *et al.*, 1997) and Russia (MESEZHNIKOV, 1988a; MELEDINA, 1994). In spite of a strong provincialism, the coexistence of ammonites, radiolarians, foraminiferas and nannoplankton allows reliable dating and correlations for the NE periphery of the Russian platform; Kimmeridgian strata which yield radiolarians are well represented in the platform environments of the Timan - Pechora basin (DE WEVER & VISHNEVSKAYA, 1997; VISHNEVSKAJA *et al.*, 1999).

The sedimentation is dominated by fine clastics (clays and silts), with limited coarser sediment supply from the remote surrounding emerged areas; marly and limy facies are very rare. Sands and silts are frequent along the North Central Ural highs and in the Pripyat basin which lies along the Fenno-Scandian shield; in the Voronezh basin, marine clastics grade into coastal plain to continental environments which develop alongside the NW border of the Stavropol high. Everywhere, the lack of subsidence and the low sediment supply suggest that the eustatic variations are the main factor controlling the sedimentation. The several formations recognised are frequently bounded by an erosional unconformity overlaid by thin coarse-grained layers, accumulations of bivalve and gastropod shells which suggest a shallowing episode at the base of the sequence; upward, condensed levels with frequent oxidised surfaces, phosphatic and pyritic nodules, glauconite and fossil accumulation witness a reduced sedimentation rate (SAHAGIAN *et al.*, 1996; HANTZPERGUE *et al.*, 1998a and b; VISHNEVSKAYA *et al.*, 1999).

In the Pripyat - Dniepr - Donetz Basin, Kimmeridgian marly-sandy to sandy-limy shallow water sequences thin smoothly westward, separated from the Lithuanian - Baltic platform - Polish basin by an area without Jurassic deposits. A reduced regional basin subsidence without conspicuous faulting may took place along the northern border of the Ukrainian shield (STOVBA *et al.*, 1996; VAN WEES *et al.*, 1996; STEPHENSON *et al.*, 2000); but, a later erosion should be responsible too (ULMISHEK *et al.*, 1994). Whatever it is, these deposits are used to emphasise a W-E seaway between Russian platform and Central to West European areas.

Radiolarian-bearing organic black shales and bituminous clay were deposited in anoxic environments. Referring to the biostratigraphic data based on ammonites, the stratigraphical distribution of the Kimmeridgian deposits of the Russian platform is reduced to the lower part of the stage (Cymodoce Zone). If compared to the main organic-rich intervals from the North Sea and West European areas of Late Kimmeridgian age, they are older. Moreover, no black shales deposits exist at that time on the central Russian platform (HANTZPERGE *et al.*, 1998b), while they may be developed in the east Middle Volga basin, if the biostratigraphic dating for the Kimmeridgian of this area is well established, regarding the revised zonal schemes.

Sea surface palaeotemperatures of the Russian platform is rather stable during the Kimmeridgian, fluctuating between 16 to 19°C, a slightly lower value than in Callovian (RIBOULLEAU *et al.*, 1998). These observations run in the same direction as data collected in more septentrional areas where Kimmeridgian samples give cooler palaeotemperatures near 9.5°C (King Karls Land, Svalbard; DITCHFIELD, 1997); they reinforce a latitudinal temperature gradient. The rise of temperatures from the Callovian to the Oxfordian, then its slight decrease and stability in Kimmeridgian correspond respectively to a global sea level rise and a maximum. Such variations are coeval with the southward increase of Boreal faunal influences on the Russian platform; the observed warming of the high latitudes in Callovian, due to the enlarged connections with the Tethys ocean, are then progressively counterbalanced by cool water coming from high latitudes.

III.2.- Turan plate

When it is possible to identify, in reduced areas and only in subsurface because it never outcrops (VOLOZH *et al.*, 1997), the Late Jurassic is everywhere dominated by varied limestones and dolostones, with coral build-ups and evaporites (South Caspian area, Kopet Dagh range and Karakum depression) or interbedded sandstones and shales (Transcaspian basin, North Ustyurt depression) deposited in shallow water marginal platforms and coastal plain environments. The Kimmeridgian and Tithonian are very difficult to separate because there is not any illustrations or descriptions of fauna from these levels, and subsequently a complete lack of real biostratigraphic data. Therefore, alternative palaeogeographic models are proposed.

On the one hand, most of the Turan domain appears without characterised Kimmeridgian, except along its western borders, all along the present day Caspian Sea; if compared to the Callovian palaeogeography, Mangyshlak, Kara Bogaz highs and several reduced uplifted areas would be emerged. But, on the other hand, it may be possible that a more extended sea covered the whole Turan domain, and deposits were subsequently eroded. As a proof, in some places, for example in the South Mangyshlak depression (GAETANI *et al.*, 1998), the Early Oxfordian dark grey-shales are capped by a regional discontinuity underlined by an ironstone condensed level, uncon-

formably overlain by a Berriasian oyster bank. In this case, the complete missing of the Late Jurassic is certainly linked to the "Neo-Cimmerian events" which may be responsible of the erosion of the most of the Late Jurassic.

Whatever would be the chosen solution, the Kimmeridgian is certainly represented in place, in continuation of the shallow platform carbonates with build ups which develop during the Oxfordian in the North Caucasus margin, Great Balkan, South Caspian and Kopet Dagh, and farther to the east up to the Amu Darya depressions. Regular eustatic fluctuations and medium subsidence rate, coupled with the absence of clastic input and a supposed dry-hot climatic regime were responsible for the accumulation of thick carbonates and evaporites until the end of the Jurassic.

III.3.- Scythian platform - Crimea - Black sea - Caucasus and Precaspian areas

The back-arc extension initiated in Callovian continues in the Early Kimmeridgian. Following the changes of configuration of the basins and the uplift of the south margin of the Scythian platform during the Oxfordian, the sedimentation has a great diversity. Carbonated deposits with build-ups extend elsewhere on platforms (Fore-Caucasus) which surround the emerged Ukrainian and Stavropol highs up to he Precaspian areas, separated from them by a narrow fringe of continental to coastal shallow marine clastic deposits. The sedimentation is mainly controlled by a quiet thermal subsidence regime, characterised by a decreasing rate (ERSHOV *et al.*, submitted). Short but active rifting phases, extension and high rate of subsidence plays as well as on the platform areas as in the South Crimea - North Black Sea belt and Great Caucasus trough. The slopes and bottom fill with thick marly deep water sediments and turbidites - flysch-like deposits, sometimes with submarine fans. The South Caspian basin undergoes too active extension and rapid subsidence. The Transcaucasus - Pontides volcanic belt is active at that time, suggesting once more a back-arc origin of the rifting (ROBINSON *et al.*, 1996; ROBINSON & KERUSOV, 1997; NIKISHIN *et al.*, 1998a and b, 2000). In the east Crimean Mounts terrane, collisional events with the Scythian platform and micro-continental blocks to the south, are recorded in the terminal Early Kimmeridgian (VOZNESENSKY *et al.*, 1998).

The Precaspian areas are still a seaway between the Russian platform and the Caucasus trough; it is narrowed compared with the Callovian because of the development of shallow carbonated platform deposits with build-ups on the eastern border of the Stavropol high. The western Precaspian depression still undergoes regular thermal cooling subsidence rates. Early Kimmeridgian is not proved in the south part of Precaspian depression, as a prolongation of the Turan platform.

An alternative palaeogeographic interpretation would be used to illustrate these domains. In South Caspian area, crustal separation would be probable since Callovian, followed by Late Jurassic sea-floor spreading (NIKISHIN *et al.*, 2000; BRUNET *et al.*, submitted). Subsequently, a narrow oceanic crust might have been drawn between the Central Iran - Alborz block rim and the South Turan rim plate, occupied by shallow carbonated platforms and slopes with siliceous-marly sediments.

III.4.- Teisseyre/Tornquist zone - Moesian platform

The palaeogeographic evolution of the Polish part of the Teisseyre - Tornquist zone and the Ukrainian Carpathians seams to be not so strongly tectonically linked to that of the Pre-Dobrogea - Dobrogea and Moesia than it was previously since the Bathonian - Callovian. As discussed for the Middle Callovian map, the opening of the wide Tethyan - Ligurian Magura areas with a narrow oceanic ridge would be coeval with successive slips of the Dobrogea faults generating an eastward displacement of Moesia. The Peceneaga - Kamena fault is henceforth considered as a key tectonic feature in the Jurassic events (HIPPOLYTE *et al.*, 1996; BANKS, 1997; BANKS & ROBINSON, 1997; TARI *et al.*, 1997). The adopted drawing is that these faults, yet hypothetically NW prolonged on the Middle Callovian map between the Polish basin and the Dobrogea - Moesia complex, would be discontinuous on the Early Kimmeridgian map. The palaeogeographic evolution is driven too by the constant sea-level rise which maximum is within the Late Kimmeridgian.

The great transgression which began in the Late Callovian - Early Oxfordian, will span all over Oxfordian, Kimmeridgian and Tithonian; in the Polish basin, it leads to communications and faunal exchanges both with the Boreal and Tethyan realms. Moreover, the opening to the Tethys and the subsequent influence of warmer sea-water caused the northward advance of carbonate platforms and the development of biohermal sponge build-ups, especially in the Oxfordian (KUTEK, 1994; MATYJA & WIERBOWSKI, 1995, 1996). Thus, as an eastern prolongation of the south margin of the West European platform, most of the Polish domain belongs to the shelf of the Tethys. At the same time, the Fenno-Scandian shield becomes again an active source of clastic material which is shed all along it, from the Danish basin to the South Baltic basin.

The carbonated facies are dominant in the SE parts of the Polish basin, however their extent toward the north is less than it was during the Oxfordian (NIEMCZYCKA & BROCHWICZ-LEWINSKI, 1988; KUTEK, 1994; NIEMCZYCKA *et al.*, 1997; DADLEZ *et al.*, 1998; KUTEK, 2000). In Pomerania and Central Poland, the facies are mainly marly limestones and marls of lagoonal to calm shallow platforms, with intercalations of sandy and dolomitic marls. Limestones are more abundant to the SE, in the Lublin area. Argillaceous marls contain molluscs assemblages with rich ammonite faunas which allow precise Early Kimmeridgian dating (MATYJA & WIERZBOWSKI, 1995; GUTOWSKI, 1998). Shallow marine sandstones, siltstones and claystones are dominant in the north and NE parts of the Polish basin as far as the Baltic platform; less abundant

ammonites and foraminifera assemblages give fairly good datings and correlations (MAREK & GRIGELIS, 1998). Different hypothesis for a tectonic control of the sedimentation are proposed. On the one hand, the Late Jurassic is time of a rifting phase (KUTEK, 1994, 2000) or an extensional episode for the central and NW part of the Polish trough (DADLEZ *et al.*, 1995; STEPHENSON *et al.*, submitted). On the other hand, Late Jurassic is an episode of overall subsidence, not associated with high thickness gradients and with no evidences for synsedimentary activity along the Polish trough bounding faults (HAKENBERG & SWIDROWSKA, 1997; LAMARCHE *et al.*, 1998; LAMARCHE, 1999). Whatever it is, the basin subsidence pattern changes completely as compared with the Middle Jurassic: the subsidence axis, parallel to the border of the East European platform (Tornquist - Teisseyre zone) shifted from a NE position to the SW border which was previously the stable part.

Encroached on the SW rim of the Ukrainian shield, a shallow marine carbonate platform develops during the Late Jurassic in the Ukrainian Carpathians (KUTEK, 1994; IZOTOVA & POPADYUK, 1996). The biostratigraphic control is not very precise because of scarce ammonites; it is mainly based on larger foraminiferas and algae. Thus, the Oxfordian, Kimmeridgian and Tithonian limits are difficult to trace. The Oxfordian strata comprise a barrier-complex of oolithic and bioclastic limestones sometimes biohermal (corals, sponges, algae, stromatoporids and bryozoans), which SW grades into shaly fore-reef open sea facies. The lagoonal deposits extend to the NE with siltstones, shales and sandstones bearing plant imprints; the facies and palaeoenvironments progressively up and laterally grade into coastal plain multicoloured shaly limestones, conglomerates and sandstones interfingered with evaporites. The palaeogeographic scheme of the Kimmeridgian is not strongly modified, showing a build up barrier which separates the open sea from the back-reef; the latter strata are a sequence of lagoonal dolomites, limestones and dolomitic limestones with interfingered anhydrites. The facies geometry and the NE-SW thickness growing of the sedimentary pile are characteristic of a narrow subsident passive margin, but there is no evidence of synsedimentary faulting.

The NW parts of North Dobrogea (GRADINARU, 1984, 1988) and Predobrogea depression (MOROZ *et al.*, 1997), up to the East Carpathians, formed a low relief uplifted area fringed by coastal plain to fluvio-deltaic and carbonate shelves with coral reef buildings, south grading into deeper marly to clayey facies. Similarly, the SE parts of North Dobrogea is drowned, as evidenced by the accumulation of pelagic limestones, without clastic input which indicates that the uplifted movements had ceased with the beginning of the Late Jurassic (SEGHEDI, 2000). However, transtensional tectonic reactivation of the Peceneaga - Kamena fault is proved by local basalt extrusions with air-fall tuffs (GRADINARU, 1988), overlaying Oxfordian – Kimmeridgian deep-water platform carbonates (SANDULESCU *et al.*, 1995; HIPPOLYTE *et al.*, 1996; SEGHEDI, 2000). The Central and South Dobrogea, probably with reduced and very low relief emerged areas, undergo a shallow marine carbonated sedimentation (GRADINARU & BARBULESCU, 1994; AVRAM *et al.*, 1995), which may be considered as the northern termination of both the Romanian part of the Moesian platform (GRADINARU, 1984) and the Bulgarian east Moesian platform (TCHOUMATCHENCO & SAPUNOV, 1994). The biostratigraphic control is mainly based on brachiopods and larger foraminiferas because ammonites are very rare.

The reconstruction of the tectonic structures of the Moesian area which takes place in Late Dogger is not fundamentally modified during the Oxfordian - Kimmeridgian, until the latest Jurassic. The sedimentation is overall conditioned by episodes of bathymetric differentiation below the influence of the successive Late Jurassic transgressions (TCHOUMATCHENCO & SAPUNOV, 1994). A very reduced tectonic activity sometimes underlines the shape and situation of the previously active horsts and grabens; the principal NW-SE fault set delimits the central Moesian basin from the west and east Moesian platform which are the site of carbonate sedimentation, here and there with build-ups which install on a ramp morphology. From NE to SW, and SW to NE, a gradation of facies from shallow-marine through deeper ramp to basinal facies makes the transition respectively between the eastern Moesian and the W Moesian platforms, with the Central Moesian basin which fills with nodular limestones and facies of *ammonitico rosso* (SAPUNOV & TCHOUMATCHENCO, 1987; SAPUNOV *et al.*, 1985, 1988, 1990; HARBURY & COHEN, 1997). In east direction, the western Moesian platform is interfingered in a complicated manner with the Central basin through the Dragoman, Vraca and Gomotarci horsts. Alongside the still emerged but reduced Thracian - Rhodopes massif, the Central basin will southward widen in the Middle Kimmeridgian, while the Nis - Trojan trough prolongs to the west in the Eastern Balkanides (TCHOUMATCHENCO *et al.*, 1989).

III.5.- Western Europe platform

The Central Atlantic (MASSON & MILES, 1984) and the Ligurian Tethys (LEMOINE & GRACIANSKY, 1988) spreading systems are active and connected through the Iberia - Africa transfer zone (VERA, 2000). The North Sea dome-wide deflation has begun since the Middle - Late Oxfordian and the rifting activity has been initiated near the Oxfordian - Kimmeridgian boundary (UNDERHILL & PARTINGTON, 1993). The transgression is everywhere evident and the marine communications between the Boreal and the Mediterranean - Tethyan realms are easy through several seaways directly crossing the W-NE and central European platform. The whole area undergoes the associated tectonic, eustatic and sedimentary effects of the "North Sea Cycle", and the Central Atlantic - Ligurian Tethys spreading which major consequence is a generalized rifting reactivation (JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998).

The sedimentation is dominated in the north by distal platform fine clastics, which the most famous and extended formation in United Kingdom and North Sea is the "Kimmeridge clays", with organic rich layers, the major hydrocarbon source rock of NW Europe. Sands

and silts of coastal marine to shallow terrigenous platform environments are restricted to the rim of several dry lands, the Fenno-Scandian shield being supposed the most elevated of these emerged areas. Shallow marine platform or ramp carbonate deposits, with frequent coral build-ups are dominant on the whole Southern Europe, northward extending as far as NW Iberia, and on the south border of the Armorican massif, the Paris basin, the Bohemian massif and the Polish trough. These facies are separated by an intermediate belt of shaly and marly facies, frequently with abundant bivalve associations and sponges, the latter sometimes making bioherms.

The onset of the rifting on the Lusitanian - Atlantic rim of Iberia is Middle to Late Oxfordian with a Late Oxfordian climax and an immediate latest Late Oxfordian - Early Kimmeridgian post-rift phase (LEINFELDER, 1993); the latter is underlined by siliciclastic inputs over much of the central part of the basin, although shallow carbonate facies and deeper marls and shales sedimentation persisted in the north and the south (WILSON, 1988). From Galicia to Algarve, the margin is subdivided into sub-basins and uplifted areas; some of them could be emergent or with restricted environments (WERNER, 1986). The Early Kimmeridgian shallowing-up sandy limestones and marls of central Portugal yield siliceous sponge facies, while in Algarve they grade into mixed sponge/coral-reef limestones; the facies succession and distribution show an E-W deepening, controlled by eustatism and tectonics (LEINFELDER, 1993; LEINFELDER *et al.*, 1993; LEINFELDER & WILSON, 1998). Controversial tectonic evolution is proposed: either halokinetic (WILSON *et al.*, 1991) or classic rifting (CANÉROT *et al.*, 1995), coinciding with the opening of the Central Atlantic and supposed to be related to a "super-plume" (WILSON, 1997).

A reactivated rifting begins during the Early Kimmeridgian in the Iberian basin of NE Spain; its effects will culminate in the Late Hauterivian. A listric extensional normal fault system structured several sub-basins at successive times; rift-induced subsidence started in Maestrat during the Early Oxfordian and in the Cameros during Early Tithonian (CANÉROT, 1989, 1991; SALAS *et al.*, 2000). As in the majority of the West Europe, the relative sea-level variations show a constant rise since the onset of the Kimmeridgian with a NW to SE deepening facies succession (BADENAS & AURELL, 1997; AURELL *et al.*, 1998, 2000). Along the emerged Iberian massif, the marginal and shallow areas of the basin are dominated by siliciclastic deposits, sometimes with reefal and bioclastic limy facies which develop onto the Late Kimmeridgian (AURELL & BÁDENAS, 1997); open shelf marls and condensed levels make the transition between the distal and the deeper parts of the basin (AURELL & MELENDEZ, 1993).

The Oxfordian to Early Kimmeridgian Palaeogeographic evolution of the Pyrenees (CANÉROT, 1989; 1991; VERGÉS & GARCÍA-SENZ, 2000) and the Aquitaine basin (BRUNET, 1984; LE VOT *et al.*, 1996) corresponds to a phase of differentiation of a platform in response to a relatively short-lived extensional-rifting phase proved by tilted blocs frame, slumps and local breccias (JAMES, 1998). Encroached on a very reduced emerged Massif Central, inner shelf deposits with evaporitic tendencies to the east, grade into more open marine environments to the west. The latter, consisting of shallowing upward carbonate cycles, yield organic-rich marine syn-rift shaly limestones, marls and shales, which constitute the main source-rocks of the Aquitaine basin.

A very similar and nearly symmetric succession of shallow evaporitic to deeper environments extends along the eastern border of the Massif Central, in the Corbières, Cévennes and Ardèche, not really strongly structured by faults as it was during the Dogger. The shallow limy facies extend along two branches which surround the deeper shaly-marly facies of the Dauphinois - Helvetic basin: on the SE, as far as the Provence - Briançonnais and Corsica - Sardinia platforms, on the NE, as far as the Jura and Schwabian - Franconian platforms. The carbonate production and clays input are sufficiently abundant to exactly balance the increase in accommodation space, resulting of constant subsidence and sea-level rises. As a consequence, shallow carbonate platform with build-ups and deeper marly facies with sponge bioherms, keep up and develop throughout the Early Kimmeridgian.

The Paris basin, the Jura platform, the Schwabian - Franconian platform and basin, the Hannover - Lower Saxony basin are permanent features, bounded by the still emerged but reduced Armorican, Central, London - Brabant, Renish and Bohemian massifs. Due to the transgression, the several marine areas are in connection, widely opened to the north by mean of corridors and troughs; to the south, they are facing the Ligurian - Magura Tethys, off a vast and quite continuous shallow carbonated platform.

In the Paris basin, bivalve-rich argillaceous limestones and marls, sometimes with biodetritic layers and build-ups are the most frequent deposits (RIOULT *et al.*, 1985; SAMSON *et al.*, 1996); they illustrate a type of prograding distal platform dominated by carbonated mud and clay deposits with build-ups, except in the Boulonnais where environments are more proximal. These facies extend homogeneously between the varied shallow water limestones of the platform which borders the north rim of the Massif Central, and the deeper clays of the Channel - Wessex basins in the United Kingdom. The tectonic control of the sedimentation still exists but the Kimmeridgian and probably the Tithonian are the moment when the subsidence speed is maximum and homogeneous all over the basin (GUILLOCHEAU *et al.*, 1999; ROBIN *et al.*, 2000).

The Schwabian - North Franconian marl basin and the Franconian limestone platform are stable structural elements persistent in time. The sponge bioherms have their main occurrence on the Franconian platform while the less homogeneous and more irregular Schwabian sponge build-ups probably developed in connection with the northern part of the Swiss coral-reef platform (MEYER & SCHMIDT-KALER, 1989; BRACHERT, 1992; SAMSON, 1997; GYGI *et al.*, 1998; ; GYGI, 2000). In the Lower Saxony (NW Germany) and in the eastern basins, uniform conditions pertained during the "Unterer

Kimmeridge" which comprises alternations of marly shales with micritic to detrital limestones deposited on a vast distal platform, divided into several sub-basins (KÖLBEL, 1968; KLASSEN, 1984; GRAMANN *et al.*, 1997). Alike during the Callovian, the existence of the "Saxonian trough", a NW-SE seaway crossing the Bohemian massif can be based on erosional remnants of Kimmeridgian deposits (MALKOVSKY, 1987). Thus, synsedimentary faulting may be inferred, related to the reactivation of extension and rifting in the Polish basin and North Sea.

As a continuation of the Callovian and Oxfordian, the palaeogeographic evolution of the northern NW European region is totally controlled by the relative sea-level variations of the "North Sea cycle" (STEEL, 1993; PARTINGTON *et al.*, 1993a and b; JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998) and by the tectonic phases of the reactivated North Sea rifting. The map locates at time when the active extension and its culmination in the Early - Mid Kimmeridgian is everywhere registered, overall in the Viking graben, north and central North Sea (CLARCK *et al.*, 1993; RATTEY & HAYWARD, 1993; LEPERCQ & GAULIER, 1996; THOMAS & COWARD, 1996). The rifting seams to be propagated following a N-S direction. From the Western Approaches - Dorset - Channel basins to the east Shetland basin and northern Viking graben, the most famous formation concerned is the "Lower Kimmeridge Clay" and its stratigraphical equivalent formations; they have been recognized in the majority of the several sub-basins intensely investigated because of their source rock capacities. Numerous detailed and synthetic papers have been published, mainly dealing with formations geometry, facies distribution and biostratigraphy, palaeo-environments, sequence stratigraphy and structural evolution. The outcrops of the Dorset coast (CALLOMON & COPE, 1995), west Weald basin, London platform, east Middland shelf (WIGNALL, 1991; WIGNALL & HALLAM, 1991) and Yorkshire - Cleveland basin (RAWSON & WRIGHT, 1995) provide the basic data, generally fairly well dated by ammonites. Offshore records are numerous, more often dated with palynology and micropalaeontology but with a lesser precision. As a whole, the most important "Oil-shale" beds development is younger (Late Kimmeridgian Eudoxus - Autissiodorensis Zones) than the time slice of the map (Early Kimmeridgian Cymodoce Zone), so they must not be reported; nevertheless, they are indicated when their dating was quoted in the referred papers as Early (or only) Kimmeridgian, or only referring to their salient lithological features, "Kimmeridge clay", with no more precision.

In addition to relevant synthesis (BROWN, 1990), the selected papers are mostly informative for the Channel basin (HAMBLIN *et al.*, 1992), Cardigan bay and Bristol channel (TAPPIN *et al.*, 1994), Moray Firth (ANDREWS *et al.*, 1990; CASEY *et al.*, 1993; STEPHEN & DAVIES, 1998), south and central North Sea grabens (HERNGREEN & WONG, 1989; CAMERON *et al.*, 1992; DONOVAN *et al.*, 1993; PRICE *et al.*, 1993; WAKEFIELD *et al.*, 1993; VAN ADRICHEM-BOOGAERT & KOUWE, 1997; GATLIFF *et al.*, 1994; KOCKEL, 1995); Norwegian North Sea – Ergesund - Danish basin (JOHANNESSEN & ANDSBJERG, 1993; STEWART, 1993) and Viking graben (CHERRY, 1993; GARLAND, 1993; MILTON, 1993; STEEL, 1993).

Generally, the "Lower Kimmeridge Clay" is composed of undifferentiated mudstones and calcareous mudstones deposited in marine low-energy environment; interbedded with silty mudstones and siltstones in the lower part, the intercalations of bituminous layers and thin limestones occur in the second half of the formation. Two divergent views have been developed to explain the depositional environments: an high productivity of algae resulting in the temporary deoxygenating of marine waters, produces anaerobic bottom sea conditions to preserve the organic matter; on the contrary, the algal blooms are the result of widespread anaerobic conditions and not their cause. More terrigenous coastal marine facies are developed on the border of emerged areas (Celtic Sea basin, Moray Firth, Horda basin). North of the emerged London - Brabant - Renish massif, fluviatile and coastal plain clastics fill in the active grabens of the West Netherlands basin which extend until the Dutch central North Sea.

III.6.- Maghreb (Morocco - Algeria - Tunisia) - Saharan areas

The "final differentiation stage" of Maghreb is reached during the Middle or Late Oxfordian onwards (ELMI, 1996). The main trends are an acceleration of subsidence (VIALLY *et al.*, 1994; ELLOUZ *et al.*, submitted), an enhanced transcurrent regime (BRACENE *et al.*, submitted), a maximum transgression which is registered at the top of the Hypselocyclum Zone or at the base of the Divisum Zone and a coeval thick sedimentation. A tilted blocks structured margin, with northward vergence (CHOTIN *et al.*, 2000), drives to the increase of deepening of the Tell (BENEST *et al.*, 1993; ATROPS & BENEST, 1994), with here and there persistent pelagic shoals and reduced deposits (ATROPS & BENEST, 1993). From Middle Oxfordian to Early Kimmeridgian, a deltaic complex, north and east bounded by a sandy strand plain with a barrier of oolitic and bioclastic limestones, settles in Rif basin and foreland (FAVRE *et al.*, 1991; CATTANEO, 1991), guided by the extension along a NE - E-NE and SW - W-SW trend, linked to a general divergence between Africa and Iberia.

Deltaic and coastal plain to coastal marine environments extend into the Atlasic (High and Middle Atlas) and Saharan domains (Lower and North Sahara basins, Oued Mya and Dahar basins), up to the Tataouine and Ghadames basins (KAMOUN *et al.*, 1999). Terrigenous input (sands and silts) is a constant and outstanding feature of the various calcareous, dolomitic and marly facies that develop in the Saharan domain, frequently with evaporitic intercalations. But, the lithology and the fossil record suggest a persistent more humid climate than during the Liassic - Early Dogger times (LEFRANC & GUIRAUD, 1990; BUSSON & CORNÉE, 1991). Coastal to shallow marine platform carbonates, sometimes with coral build-ups, and marly limestones are dominant from the High Atlas to the Chotts area and Gulf of Gabès (BELHAJ, 1996), through the Lower

Sahara basin. In the Middle Atlas, the northern fringe of the deltaic series is drowned by the major Early Kimmeridgian and onwards sea level-rise, driving to the edification of a part of the Late Jurassic carbonate platform which henceforth extends from the Oran High Plains to the Tlemcen - Preatlasic domains.

Deeper and pelagic terrigenous facies (marly limestones and marls) are restricted to a narrow fringe North to the Middle Atlas while they widely develop in the Atlasic Tunisia, North-South Axis and Tunisian Dorsale, but without or very weak and fine clastic input (PEYBERNÈS *et al.*, 1990; PEYBERNÈS, 1992; PEYBERNÈS *et al.*, 1996; SOUSSI *et al.*, 1991a and b, 1999, 2000). Plants associations of the fluviatile to proximal marine environments of the Dahar basin demonstrate the persistence of rapid development of successive soils, episodically drowned by transgressive sand levels (BARALE *et al.*, 1998). A N-S active extension and a moderate thermal subsidence rates characterise the Tunisian areas till the latest Jurassic (BARRIER *et al.*, 1993; HLAIEM *et al.*, 1997; BOUAZIZ, 1995; BOUAZIZ *et al.*, 1994, 1996, 1998, 1999; PATRIAT *et al.*, submitted).

Late Jurassic data are scarce in the Essaouira-Agadir basin; the Oxfordian, Kimmeridgian and Tithonian are often difficult to separate (FAVRE & STAMPFLI, 1992; BROUGHTON & TRÉPANIER, 1993; MEDINA, 1994; MORABET *et al.*, 1998). A high subsidence rate is compensated by an abundant terrigenous input provided by the Saharan craton, the Moroccan Meseta and the henceforth emerged western High Atlas. Generally, inland outcrops and boreholes show detrital facies (clays, silts and sands) with frequent carbonaceous, dolomitic and evaporitic intercalations of coastal plain to transitional environments. Deltaic red beds are recorded from the Argana areas (MEDINA, 1994) and marine platform carbonates to deeper carbonates in offshore boreholes. The displacement of the shore-line of the Essaouira - Agadir gulf reaches a maximum towards the east and the north, off El-Jadida, as a result of the major Early Kimmeridgian transgression and the beginning of the major Late Jurassic - Early Cretaceous subsidence phase (MEDINA, 1994, 1995; LABBASSI *et al.*, 2000; CHOTIN *et al.*, 2000).

The coeval Tethys and Atlantic spreadings during the Late Jurassic (Oxfordian - Kimmeridgian) induce a deepening of the Maghreb Tethyan margin and its Atlantic façade which seems to run counter the Atlasic spreading evolution, initiated since the Early Liassic.

III.7.- Egypt - Sudan - Libya

The Kimmeridgian major transgression induce a SW retrogradation of the coastal marine deposits and the widespread of continental - fluviatile environments on the several previously erosional areas (SCHANDELMEIER *et al.*, 1997; GUIRAUD *et al.*, 2000). A continuous belt of parallel and mixed shallow carbonated-clastic platform overlaps the present day Mediterranean coast line, inland bordered by coastal plain deposits still dominated by clastics; terrigenous input continues into sabkhas with evaporitic facies on both sides of the Farafra - Bahariya high. The pure carbonated deposits are restricted to discontinuous narrow fringes of the marine shelf, east and west of the Nile Delta and the NW Cyrenaica.

Alike in Callovian, the detailed stratigraphy and age of the several formations are still obscure because of a subsurface knowledge with rare reliable biostratigraphic data, facies upgrading and interfingering controlled by sea-level variations and regional tectonic events (ABDEL AAL *et al.*, 1990; HANTAR, 1990; JENKINS, 1990; KERDANY & CHERIF, 1990). The Kimmeridgian age is mainly deduced from palynological and foraminiferas associations. The Upper part of shallow sandy-silty carbonates in Sinai is dated Kimmeridgian; it should upgrade into the Tithonian on the base of large foraminifera and calcareous algae (JENKINS, 1990; KERDANY & CHERIF, 1990). Palynomorphs, foraminifera and calcareous algae have provided a Kimmeridgian age for the shallow marine carbonates and the mixed facies from the subsurface of the Nile Delta (KEELEY *et al.*, 1990). Clastic rich limestones, dolomitic limestones, shales and marls from the NE Libyan coast and offshore Cyrenaica, are dated as Kimmeridgian to Neocomian with miospores and dinocysts (THUSU *et al.*, 1988; DURONIO *et al.*, 1991); the facies laterally intergrade from tidal flat complex-fluvial clastics to intertidal - inner neritic, prodelta - outer neritic and deeper fine clastics. Fluvial to sand-flat and intertidal clastics rich deposits of the Galala plateau (north Gulf of Suez), ranges from late Middle Jurassic (?Callovian) to Late Jurassic (Kimmeridgian - ?Tithonian) as indicated by the sequence stratigraphy interpretation (DARWISH, 1992); these Late Jurassic formations unconformably overly Late Triassic or Late Carboniferous - Permian facies with definite truncation and erosion surface. Here and there, in the shallow marine to coastal plain - sabkhas environments, some black shales and coal rich layers are interbedded with sandy or dolomitic limestones.

The Dahkla, Erdis and Al Kufrah basins still fill with continental-fluviatile clastics; they are in connection and enlarged with regard to the Middle Jurassic. The Kimmeridgian may be registered (KLITZSCH & WYCSISK, 1987; HERMINA, 1990; KLITZSCH, 1990; KLITZSCH & SQUYRES, 1990; KLITZSCH & SCHANDELMEIER, 1990). It is expected that large areas of the central and southern Sirt basin, especially the deeper parts, yield Late Jurassic "Nubian sandstones" facies which covers the carboniferous basement (EL-HAWAT, 1992, EL-HAWAT *et al.*, 1996; WENNEKERS *et al.*, 1996); but, a connection with the eastern Libya and Western desert is more probable than a direct opening on the Gulf of Sirt.

The sedimentation is still guided by both a SW-NE wrenching (KEELEY & WALLIS, 1991; ANKETELL, 1996; EL- HAWAT, 1992; EL-HAWAT *et al.*, 1996; SMITH & KARKI, 1996; KEELEY & MASSOUD, 1998; AYYAD *et al.*, 1998) and a W-E normal faulting (MOUSTAFA & KHALIL, 1990; MOUSTAFA *et al.*, 1998).

III.8.- Levant (Israel - Lebanon - Syria - Jordan)

The Early Kimmeridgian is developed in Israel in a shallow marine carbonated facies with reefs in central Judea until North Negev and North Sinai, oolites in

Galilee, everywhere interbedded with marls, shales and sands (Hirsch *et al.*, 1998). Several of these facies are supposed to extend up to South Negev and Sinai; clastic input is more marked in South Negev. During Liassic and Dogger, clastics came from the wearing down of the Arabian shield in the south and south-east. Along the present day Levantine offshore, yet in the Early Oxfordian and until the Kimmeridgian, clastics input increases westward; this suggests a derivation from an emerged area now probably included within the composite Tethyan platform of Erathosten - Rhodos - Bay Daglari (Hirsch *et al.*, 1995). Large parts of the Late Jurassic deposits were eroded during the infra-cretaceous uplift along the present day Levantine coast (Gevar'am canyon), in the Negev and Judea up to Jordan, making the delimitation of the shoreline very hypothetical.

The Late Jurassic formations of Lebanon are diversified, with shales and marls, marly limestones and limestones, dated Oxfordian in there lower part. Basalts intrusions and pyroclastics are interbedded in Mount Lebanon, linked to a block faulting phase associated with a gentle uplift and regressive to emergent trends (Walley, 1997). Following the tectonic quiescence which span since the Early Liassic, the Oxfordian volcanic "Bahnnes event" can be correlated with the Devorah volcanism in Israel, both marking the end of the stability over the Levantine area (Laws & Wilson, 1997). The Early (and possibly Mid) Kimmeridgian can be palaeontologicaly recognised within the upper part of the sediments, up to the volcanics intrusions. The associated facies indicate a transgressive episode from mid shelf to near shore environments in the overlying Middle to Late Kimmeridgian formations.

In the interior Syria (Palmyrides) as far as West Iraq (west Rutbah high border) and Jordan (Arad high), the Late Bathonian - Callovian regressive phase progressively withdraws the Jurassic sea from the platforms and rifted basins to the Tethyan Ocean in the north and to the east Mediterranean basin in the west (Mouty, 1997a). The resulting Early Kimmeridgian platform, thus Late Kimmeridgian embayment on the Syrian Coastal chain, continues to gently subside until the end of the stage, as do the Lebanon Mount, Anti-Lebanon and Galilee. Despite a well enough dating by large foraminiferas (Mouty, 1997b), the limy-marly formations of the Syrian Coastal chain are still unprecisely correlated with the formations of the Galilee - Lebanon areas. As a consequence, the "Bahnnes" pyroclastics and basalts intrusions should be either latest Oxfordian or earliest Kimmeridgian (Mouty *et al.*, 1992; Walley, 1997).

Kimmeridgian is absent in the Sinjar trough and the Euphrates graben; dolomitic to marly limestones in the Mesopotamian foldbelt are doubtfully considered as partly Kimmeridgian (Alsharhan & Nairn, 1997).

Western Jordan is still an emerged area and northern Jordan a coastline where continental to restricted and nearshore varied deposits onlap the northern flank of the Rutbah High (Bandel, 1981; Alsharhan & Nairn, 1997).

III.9.- Central Arabian platform and Gulf area - Iraqi platform - Oman - Zagros basin

During the Early Kimmeridgian, the Arabian plate appears as a huge shallow carbonated ramp, widely east opened on the Tethys ocean and west underlined by a narrow siliciclastic coastal fringe (Murris, 1980). From the emerged Arabian shield, the eastward succession of palaeoenvironments goes from alluvial plain, lower coastal plain, shallow clastic, mixed and carbonate to evaporite platform, onto the Zagros foldbelt and the Tethyan passive margin isolated platforms. On the south, an emerged barrier between Dhofar High and the arabo-nubian shield, prevents connection with Yemen and the "African Horn"; on the north, the Iraqi - Iranian platform, including the Zagros platform, is in connection with the Levantine areas, south of the still emerged Mardin High.

In central Saudi Arabia, Early Kimmeridgian deposits exist within formations, some of them being the main Jurassic source rock in the region. The precise age of these limy shallow marine deposits is difficult to establish in spite of abundant large foraminiferas (Hughes, 1996; Le Nindre *et al.*, 1997) and some but more often endemic ammonites (Enay & Mangold, 1994). Sequence stratigraphy approach is thus an example which drives to the best possible but still confusing correlations due to variable nomenclature and scarce biostratigraphic data: the culmination of an Early Kimmeridgian transgressive cycle overlain by a regressive sequence is registered (Le Nindre *et al.*, 1990b).

Therefore, reliable correlations are proposed between Upper Jurassic units of several areas of the Arabian plate (Grabowsky & Norton, 1995; Alsharhan & Nairn, 1997; Al Husseini, 1997) which yield, within a mostly limy succession, a carbonaceous shaly episode more or less bituminous in Qatar and the United Arab Emirates (De Matos & Hulstrand, 1995; Al Silwadi *et al.*, 1996) and Kuwait (Youssif & Nouman, 1995; Carman, 1996; Hughes, 1996). Similar facies in North Iraq and in Zagros would be too good candidates for such correlations.

A thick limy equivalent of the central Saudi Arabia formation is recognised within shallow marine carbonate deposition in the Oman Mountains and Interior Oman, while to the north it would be a condensed carbonated sedimentation in the Musandam Peninsula (Le Métour *et al.*, 1995).

In Saudi Arabia and Oman, the Kimmeridgian is characterised by a low rate of sedimentation, mainly controlled by eustatism during a long term rise in sea level which began in Oxfordian until the Early Kimmeridgian highstand. However, alike during the Middle - Late Callovian, normal faults system in the western gulf area (Carman, 1996), mainly orientated in a parallel direction to the passive Arabian plate margin, enhanced the resulting sedimentation. The factors controlling the sedimentation are both a gentle regional intraplate stress which induces a reactivation of subsidence, coupled with a sea level rise. In the south, the

beginning of the thermal doming activity linked to the active rifting of the Indian Ocean, certainly induces the uplift of Dhofar High and westward emersions, as well as SW-NE normal faulting in Oman.

III.10.- Ethiopia - Somalia - Gulf of Aden - Yemen

Since the onset of the main Callovian transgression, the palaeogeographic features undergo few modifications on East Africa except an E-W narrow emerged area which extends from Dhofar High in south Arabian plate up to the Nubian craton. Yemen and Gulf of Aden have once more no marine connection with Central Arabia, and the Indian Ocean definitely separates East Africa and India.

Predominantly open marine sedimentation on outer ramp represents the highstand of the Jurassic transgression which is registered in marls and marly limestones facies of central Ethiopia (FICCARELLI *et al.*, 1975; TURI, 1980; BOSELLINI *et al.*, 1997; MARTIRE *et al.*, 1998, 1999) or their equivalents on the continental margin of Somalia (BOSELLINI, 1989; LUGER *et al.*, 1994a), in the Danakil Alps (SAGRI *et al.*, 1998), in Yemen (BEYDOUN, 1997; BEYDOUN *et al.*, 1996; AL-THOUR, 1997) and the Gulf of Aden (BEYDOUN, 1989). It is supposed that Early Kimmeridgian exists in these thick sedimentary pile, in spite of scarce biostratigraphic data; Oxfordian to Middle - Late(?) Kimmeridgian is proved by various fauna (ostracods, brachiopods, belemnites and ammonites; METTE, 1993). Radiolarians, rare benthic foraminifera and nannofossil assemblages indicate a Late Oxfordian - Early Kimmeridgian age in the Danakil Alps (SAGRI *et al.*, 1998); the ammonites which are found in the upper unit of the Antalo limestone of the Mekele outlier (Tigrai) confirm the Early Kimmeridgian age (MARTIRE *et al.*, 1998, 1999).

Alike in the lower part of the Antalo limestone and equivalents, the overall depositional setting is a ramp, gently dipping to the East from inner to outer through mid environments (SAGRI *et al.*, 1998), basically controlled by tide and storm processes. Shallower carbonate deposits underline the western part of Ethiopia onto Yemen; mixed terrigenous and limy coastal marine shallow environments with fluctuating salinity install in the Abbay River-Blue Nile basin (GETANEH, 1991). In the Marib - Jawf basin and in half-grabens of the Hadramaut area in Yemen, several organic-rich shales are intercalated in limestones and distal sandy-silty marls.

Alkaline magmatic intrusions, dated as Late Jurassic in the Nuba Mountains and the Bayuda Desert in Sudan (VAIL, 1990), may be reported either on the Kimmeridgian or the Tithonian map.

The Kimmeridgian of North East Africa and Yemen provides an example of the relationship between plate margin and intraplate deformation as a consequence of the break up of India and the oceanic ridge activity. Strike-slip movement and development of transtensional basins are much more marked than in Late Middle Jurassic in the several basins affected (Blue Nile, Abbay, Ogaden, Berbera Borama, Ahl Mado, Gulf of Aden, Marib Jawf). The factors controlling the thick sedimentation should be a regional intraplate stress which induces a reactivation of subsidence (syntectonic deposition within subbasins) coupled with sea level changes.

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11.- EARLY TITHONIAN (141 - 139 Ma)

J. THIERRY[1]

I.- MAIN FEATURES

The Early Tithonian map of the Peri-Tethys Programme is a new one considering that a Late Tithonian map was included in the Tethys programme (FOURCADE *et al.*, 1993).

As well as possible, the biostratigraphic interval concerned corresponds to the Elegans Subboreal Ammonite Zone or its equivalent, the Hybonotum/ Lithographicum Submediterranean - Tethyan Zones and the Klimovi Boreal Zone.

According to the International Commission on Stratigraphy, "Tithonian" must be henceforth used as a formal name for the last Jurassic stage. "Volgian" can be used in boreal and subboreal realms as an equivalent of the Tithonian, as far as reliable correlations are established with the submediterranean and Tethyan realms. The "Kimmeridgian" (*sensu anglico*) and the "Portlandian" would not be so long used as stage names. Regarding the definition of the Kimmeridgian, Tithonian and Volgian stages (MESEZHNIKOV, 1988b; GEYSSANT, 1997; ROSTOVTSEV & PROZOROVSKY, 1997; HANTZPERGUE *et al.*, 1997, 1998a and b), the Early Tithonian of the Tethyan - Submediterranean realm corresponds to the earliest Late Kimmeridgian (*sensu anglico*; Subboreal realm), and to the earliest Early Volgian (Boreal realm).

The choice of the Early Tithonian allows better correlations than the Late Tithonian between the several realms and areas concerned. The previously chosen Durangites Zone, which takes place during a strong regressive episode, does not allow precise correlations between the marine and continental/shallow brackish facies identified and widely developed in North Western Europe (Early Portlandian and "Purbeckian" facies in England, France, Switzerland, Germany).

On the contrary, within an eustatic framework, the Early Tithonian map takes place immediately after the major Late Jurassic first order peak transgression, located near the Kimmeridgian - Tithonian boundary, which precedes the great Late Jurassic - Early Cretaceous regression in Western Europe (HARDENBOL *et al.*, 1998).

The palinspastic reconstruction used is that of the previous published Tethys map (RICOU, 1996) and based on the palaeomagnetic Early Tithonian anomalies M21/M22-M21/M18 which embraces an interval comprised between 149-143 Ma (GRADSTEIN *et al.*, 1994; GALBRUN, 1995), with a confidence interval of ± 3 Ma.

There is no available radiometric data correlated with a Tithonian ammonite zonal scheme. The closest data is that located at the Early/Middle Oxfordian boundary (FISHER & GYGI, 1989).

The polarity sequence of the Kimmeridgian - Tithonian deposits of the Subbetic province and the M-sequence of marine magnetic anomalies have a good similarity (OGG *et al.*, 1984). Coupled with precise biostratigraphic data, the revised following ages are proposed for the M-sequence (GRADSTEIN *et al.*, 1994; GALBRUN, 1995): the Kimmeridgian/Tithonian boundary is correlated with the end of the M-22A anomaly; the Early/Middle Tithonian boundary falls within the M-22 anomaly and the Middle/Late Tithonian boundary correlates with the begining of the normal polarity of the M-20 anomaly.

Noticeable divergence appears when comparing the Tithonian magnetic polarity successions, the magnetochrons and the corresponding radiometric ages. Regarding both to the GRADSTEIN's time scale (1994) and the GALBRUN's magnetic succession (1995), the Elegans - Hybonotum/Lithographicum Zones interval would be bracketted between 150.7 and 149 Ma, with an uncertainty of ± 3 Ma.

On the ODIN's time scale (1994), the lower boundary of the Tithonian, not precisely documented, is located at 141 Ma, while the upper boundary, likewise poorly documented, is at 135 Ma.; but, it would be 5 Ma older or younger.

Once again, considering a composite ammonite zonation with an equal duration assigned to each biostratigraphic unit and using the ODIN's time scale, the concerned time slice for the Early Tithonian Peri-Tethys map may be close to 141-140 Ma. with an unknown uncertainty; considering the GRADSTEIN's scale, the same biochronologic interval is much older, comprised between 147.7-143.7 Ma. When reliable biostratigraphic data are missing, such a difference would be kept in mind for any estimation or precise location of events near the Kimmeridgian/Tithonian boundary which does not overlap in the available time scales. Finally, keeping in mind all

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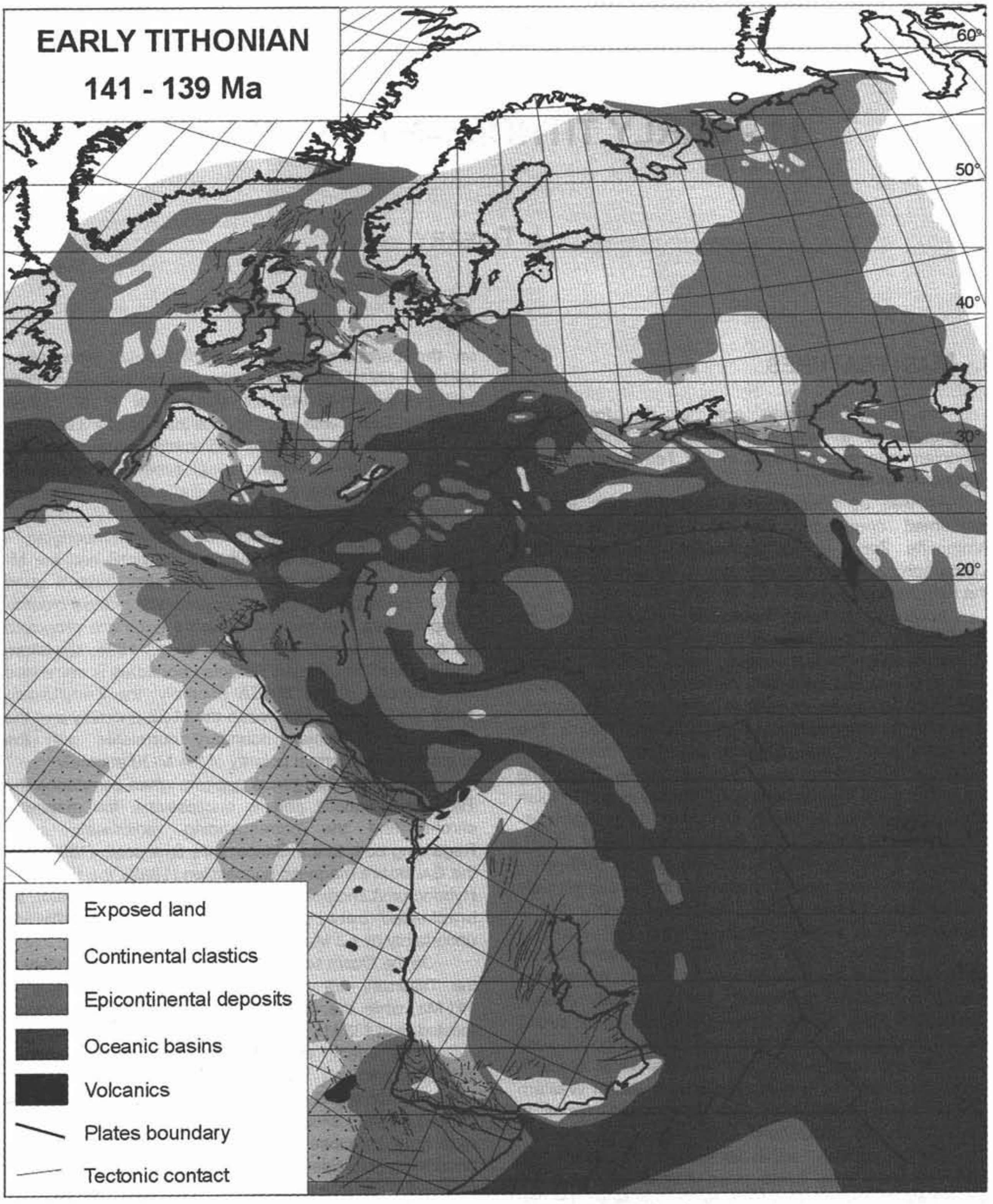

Fig. 11.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Tithonian.

the above discussed data, but giving the preference to the ODIN's time scale, the Early Tithonian map is assumed to be located between 141-139 Ma.

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for

Laurasia and West Gondwana plates keep on a gentle anticlockwise rotation which consequences remain a respectively southward and northward shifting displacement of their east and west parts; but, the Tithonian can be considered as the onset of a plate reorganisation which will be emphasised all over the Cretaceous:

1.- the Transcaspian - Caucasus - Crimea - Dobrogea alignment, beyond the north "Central Tethys" margin, continues to be controlled by the subduction of the Tethys oceanic crust which underlines the southern border of the Iranian block, the Pontides platform and the Moesian block; but, island arc activities are reduced to the Sanandaj area, along the oceanic rim of the Central Iranian block;

2.- the Atlantic has definitively acquired its typical symmetric corridor shape with oceanic crust on both sides of an active mid-oceanic ridge henceforth producing a NW-SE expansion between North America and Africa. In this way, the "Atlantic Tethys" (Central Atlantic) strongly differs from the "Central Tethys"; the latter, including the eastern oceanic area, widely opened and located between Arabia and Central Iran, is an asymmetrical ocean with northward converging and southward diverging boundaries. Since the beginning of the Jurassic, the consequences of the Atlantic opening were transmitted eastward to the Ligurian ocean and the complex triple junction point in Apulia, through the Maghrebian transfer zone. During the Late Tithonian - Early Berriasian interval, Iberia is going to separate from Laurasia and its kinematics will be better connected with North Africa displacement. Moreover, the onset of the Dinaric - Hellenic obduction and collision on the Apulian plate is a new and major tectonic element of the Central Tethys. The outstanding consequences are that the Ligurian ocean stops spreading during the Tithonian and that the complex "Mediterranean *Seuil*" is evolving toward a new frame;

3.- in the SW Central Tethys ("Mediterranean *Seuil*" or "Mediterranean Tethys") the complex palaeogeography consisting of a shallow carbonate platform mosaic and emerged areas is slightly modified by a different development of the deep marine and furrows. In spite of the Dinaric - Hellenic obduction, the propagation of Tethyan branches is still active in the Pindos - Olonos furrow supposed to be occupied by continuous oceanic crust. A westward prolongation of the Pamphylian basin is going to separate the SW Central Tethys and Apulian platforms from the Sicily - Malta escarpment and the Pelagian shelf.

4. - West Gondwana (Arabia - Somalia) and East Gondwana (India - Madagascar) are definitively separated by a continuously widening deep marine corridor: the future Indian Ocean, connected to the Indian-Malagasy basins. The eastern border of the Arabian craton remains a passive margin with a wide evaporitic platform facing the Tethys ocean. During the Late Jurassic - Early Cretaceous, an accelerated rifting activity is affecting the SE part of the Africa - Arabia cratonic areas, as evidenced by the developments of rifts in Yemen and Somalia.

The several blocks which are accounted for in the Early Tithonian are directly inherited from the Kimmeridgian palaeogeography:

1.- the East European and the Russian platforms remain the most stable parts of Laurasia; because of regressive trends and uplifting tendencies, the Tithonian - Volgian seas have a slightly reduced extension, witness of the first manifestation of the end-Jurassic general regression. The marine seaway between the East European and the Russian platforms, which previously separated the Fenno-Scandian shield from the Ukrainian high, seams to be intermittent or closed;

2.- slightly restricted marine connections still exist between the Precaspian basin - Turan plate and the Russian platform which is southward opened on the Great Caucasus basin. The Mangyshlak and Kara-Bogaz areas are part of a wide complex of shallow shelves with coastal marine to fluviatile deposits and dry lands, as the westward prolongations of the Kazakhstan plateau and the south Ural highs;

3.- the termination of the Great Caucasus marine basin seams to be not far to the east between the Iranian block and the Turan plate. The Lesser Caucasus - Pontides platforms and small emerged areas underline the henceforth narrow Great Caucasus - South Crimea basin which still undergoes active extension and fills with deep marine marly facies. All along the Scythian platform rim, the low relief and coastal areas, the shallow platforms and troughs are undergoing a regression, illustrated by restricted - evaporitic coastal and fluviatile facies;

4.- the Moesian block and its surroundings remain totally below marine conditions. Coupled with regressive tendencies, compression events and strike-slip faults activity in Crimea, in South Dobrogea and may be in West Pontides and Rhodope - Thracian massif, lead to uplifts and emerged areas. To the west, the opening of the Carpathians - Magura basin is still active but reduced and supposed without prolongation into the Ligurian basin;

5.- the SW borders of the Ukrainian and Fenno-Scandian shields undergo too a regression in spite of a thermal subsidence phase in the Polish Trough and a reactivated rifting episode in the Scania - Danish basins, probably related to the North Sea rifting events. The regression narrows the marine areas of the Polish Trough, east Carpathians and Baltic basins underlined by a huge emerged Bohemian massif. Marine connections with the North Sea are interrupted by the accumulation of fluviatile and coastal deposits;

6.- the Mid and Central Europe are similarly affected by regressive tendencies probably coupled with uplifting activities which progressively enlarge the dry lands (Rhenish, Central and Armorican massifs) and drastically reduce the marine areas (Hanover basin, Schwabian and Franconian platforms), which connections with the North Sea are interrupted. The only remaining marine connections are through the Mediterranean - Aquitaine - Pyrenees - Alpine platform and basins and the Paris basin, being opened on the United Kingdom - North Sea basins still affected by the continued dome-wide deflation associated with rifting phases;

7.- the northern rim of Iberia undergoes a relevant regression illustrated by the wide extension of shallow marine, coastal to restricted and fluviatile deposits which invade the Iberian platform. Marine connections with West Europe remain only along the tectonic alignments of the future Bay of Biscay rift which onset may occur near the Late Tithonian/Early Berriasian boundary. Extensional regime is registered on the Atlantic Lusitanian passive margin and on its Mediterranean border, facing the active Maghrebian transfer zone;

8.- the Atlasic domain, from Morocco to Tunisia, is still a tilted blocks structured margin which undergoes active rifting conditions. The whole area is marked by regressive tendencies and a general N-E shifting of the shallow marine facies while the deeper ones are developed all along the active transfer zone facing the Iberian block. The Saharan areas are still occupied by continental basins but with none or reduced evaporitic conditions;

9 - the present day shorelines of Libya and Egypt remain too a passive margin with tectonic activities; regressive conditions narrow the shallow shelf which underlines the large continental basins with none or reduced evaporitic conditions;

10.- in Levant, the development of volcanic fields seams to be more plume-related than rift-related; at the same time, an intense uplifting regime leads to a marked regression and a narrow shallow shelf;

11.- the Arabian platform is occupied by the more widespread evaporite platform than earlier in the Jurassic. The sedimentation is mainly controlled by eustatic variations; the impact of the Indian Ocean opening seams to be weak, restricted to intra-plate tectonic producing N-S orientated grabens in central Arabia;

12.- the African corner is affected by an intense rifting activity in Yemen and Somalia where the sedimentation is both controlled by eustatic variations and tectonic activities.

II.2.- Palaeoposition of plates and blocks

There is no new available data nor computed position of blocks; therefore, the palaeolatitude grid remains unchanged (BESSE & COURTILLOT, 1991). The palaeoposition of plates is that adopted for the Late Tithonian Tethys map (RICOU, 1996) where the Atlantic ocean width and the respective position of North America and Africa are constrained by the M21 anomaly (KLITGORD & SCHOUTEN, 1986). Alike for the Callovian and Kimmeridgian Peri-Tethys maps, and opposite to the Late Tithonian Tethys map (FOURCADE *et al.*, 1993), oceanic crust without sediments has been added where drifting areas would exist (Central Atlantic, Iberian -Maghrebian - south Ligurian transfer zone, Magura and Pindos - Olonos furrows). Within the sequence of the successive Jurassic palinspastic reconstructions, the position of Iberia and Corsica - Sardinia blocks are respectively W and NW shifted according to kinematics model, tectonic trends and facies distribution (FOURCADE *et al.*, 1977; CANÉROT, 1991; OLIVET, 1996; VIALLY & TRÉMOLIÈRES, 1996; VERGES & GARCIA-SENZ, 2000; VERA, 2000). Moesia has an unchanged position regarding the Tethys map (FOURCADE *et al.*, 1993).

II.3.- Accuracy

On its Atlantic margin, the position of West Gondwana is constrained by the fit of North America and Africa based on the M21 Atlantic anomaly; on its Indian margin, the position of East Gondwana is based on a postulated symmetrical expansion since the fitted time as the Mozambique anomalies M22-M16 and M22-M18 have no counterpart (RICOU, 1996). The first oceanic crust has been dated as Early Tithonian (146 ± 6 Ma) and rifting occurred at this time in southern Yemen as the Gulf of Aden - Somalia rift system develops (ELLIS *et al.*, 1996). Therefore, the reconstruction refers to the classical Atlantic fit. But, as it has been previously discussed, one must keep in mind the discrepancies which exist between the proposed time scales and the palaeomagnetic calibration (GRADSTEIN *et al.*, 1994, 1995; ODIN, 1994; GALBRUN, 1995).

Biostratigraphic datings and correlations have made great strides, especially since the construction of the Tethys maps (FOURCADE *et al.*, 1993). Very precise ammonite zonal schemes and correlations are henceforth possible, on the one hand between Tethyan - Mediterranean and Boreal faunas, and on the other hand between West European (Euro-Boreal - Mediterranean) and Russian platform faunas (HANTZPERGUE *et al.*, 1997, 1998 a & b). Several microfossil groups can be currently used in parallel scales, overall in NW Europe, including the North Sea domain and the Russian platform: radiolarians (VISHNEVSKAYA, 1995; DE WEVER & VISHNEVSKAYA, 1997; VISHNEVSKAYA *et al.*, 1999), calcareous nannofossils (RIDING & IOANNIDES, 1996; GARDIN, 1997) and dinoflagellates (DE KAENEL *et al.*, 1996; FAUCONNIER, 1997). It is called to mind that deposits containing calpionellids are not concerned by the Early Tithonian map; these microfossils give precise biostratigraphic data, correlated with ammonite zonal schemes, for younger intervals, not before the Early/Late Tithonian boundary.

Sequence stratigraphy may provide a first-step reliable correlations. The most complete succession and the hierarchy of the several order sequences and cycles for the Tithonian is based on data from NW Europe sediments (PARTINGTON *et al.*, 1993a and b; RATTEY & HAYWARD, 1993; STEEL, 1993; COE, 1996; JAMES, 1998; GUILLOCHEAU *et al.*, 1999; ROBIN *et al.*, 2000; AURELL *et al.*, 2000). Tentative correlations have been proposed with the Russian platform (SAHAGIAN *et al.*, 1996) and the Arabian Gulf (AL-HUSSEINI, 1997). Some of these data are henceforth compiled on the Jurassic sequence Chronostratigraphy / Biochronostratigraphy Chart (HARDENBOL *et al.*, 1998), but, some discrepancies still exist between successions of the Euro-Boreal realm and the Mediterranean - Tethyan realm. Referring to such a chart, the selected Early Tithonian interval falls during one of the last and best correlated "3rd order sea-level rise" which follows the Late Kimmeridgian "1st order" major peak transgression.

Finally, the stratigraphic nomenclature of the Late Jurassic subdivisions could appear confusing to somebody; thus, it is necessary to refer to official decisions (MESEZHNIKOV, 1988a and b; GEYSSANT, 1997; ROSTOVTSEV & PROZOROVSKY, 1997; HANTZPERGUE *et al.*, 1997; 1998a and b) to avoid any confusion between obsolete and recommended chronostratigraphic units. When lithologic and faciologic data are not, or imprecisely constrained by a biostratigraphic record, a comprehensive interpretation is reported, as far as possible representative of the selected interval (Early Tithonian;

= earliest Early Volgian; = earliest Late Kimmeridgian *sensu anglico*).

II.4.- General comments

During the latest Jurassic, the sinistral rotational-translation of Eurasia relative to Gondwana, associated with the successive opening phases in west central Tethys and Central Atlantic emphasises the tectonic regime governing the evolution of the Peri-Tethyan platforms (ZIEGLER *et al.*, 2000); but, the maximum activities take place during the Late Tithonian or early - Middle Berriasian, underlined in NW Europe by the so-called "Neo-Cimmerian unconformity". The evolution of the southern margin of the East European craton is still concerned by the repeated back-arc extension and compression phases which are related to the activity of the north-dipping Tethys subduction; along the southern border of the Scythian platform, Early Tithonian corresponds to the onset of compression episodes which will progressively lead from Late Tithonian to Middle Berriasian to thrusting and folding in the Dobrogea - Crimea - Black Sea alignment and to a remnant basin in Great Caucasus - Transcaspian (NIKISHIN *et al.*, 2000). The orthogonal opening and sea-floor spreading of the Central Atlantic, and the transtensional opening of the Atlantic - Alpine Tethys lead to develop a shear-like South Iberian margin (VERA, 2000); but, the sinistral displacement of the Iberian microplate with respect to the continental Europe has not yet begun, as well as its consequences: the development of a system of pull-apart basins on the Iberian platform and in the Aquitaine basin along the future Bay of Biscay rift (LE VOT *et al.*, 1996; SALAS *et al.*, 2000; VERGÈS & GARCÍA-SENZ, 2000); the opening of the oceanic Valais trough; the propagation of rifting through the North Atlantic. This will occur not before the Tithonian - Berriasian boundary, or later during the Early Berriasian.

Major eustatic events are coeval with the tectonic evolution. The Early Tithonian map locates at the onset of the most important "1st order regressive cycle" of the whole Jurassic which is the last episode of the "North Sea cycle" (JACQUIN *et al.*, 1998). The maximum regression falls in the Late Berriasian, but the several "2nd order facies cycles" and "3rd order sequences" which are recognised demonstrate a continuous shoaling of the whole northern Peri-Tethyan shelves, prior to the quite general Late Tithonian emersion. The correlations with the southern Tethyan shelves sequences are not easy because of the discrepancies between biostratigraphic data. Moreover, the resulting sequence stratigraphy framework of Laurasia seams to be better constrained by tectonics than eustasy, while on West Gondwana, tectonics seams to play a more discrete part (LE NINDRE *et al.*, 1990b, submitted; AL-HUSSEINI, 1997).

The increasing endemism and the decrease of ammonite faunal exchanges is a relevant palaeo-biogeographic feature of the Tithonian; the direct consequence is the difficult correlations between the Boreal ("Early Volgian"; Russian platform), subboreal ("Late Kimmeridgian" *sensu anglico*; NW Europe), submediterranean (Early Tithonian; SW and Central Europe) and Tethyan realms. The earliest Early Tithonian interval (Klimovi - Sokolovi / Hybonotum - Mucronatum - Vimineus - Palatinus Zones) corresponds to the moment of the last Jurassic ammonites exchanges between the Russian platform and the Submediterranean realm (HANTZPERGUE *et al.*, 1998a and b); at the same time, there is no exchange between the Russian platform and the subboreal realm. Intermittent connections are proved later between the Russian platform and the subboreal realm during the Middle Volgian and the "Portlandian" (*sensu anglico*). The regression on Laurasia and the subsequent closure of previous seaways is partly responsible for the setting of endemic faunas, as well as the development of huge shallow shelves unfavourable to ammonites ecology. The Tithonian ammonite assem-blages were less diverse in the shelves and different from those of the pelagic areas where the dispersion is better realized (CECCA, 1999).

The geographical extent of pelagic carbonates with cherts ("Maiolica" facies) and abundant microfossils (calpionellids, nannoconids), is a relevant sedimentary feature of the slopes and deep basins of the Tithonian Tethys (DE WEVER *et al.*, 1996; FOURCADE *et al.*, 1993, 1996). This facies spreads over the Atlantic and Apulian areas, replacing the shales and overall the radiolarites which formerly developed in these palaeoenvironments; at the same time, ammonitico rosso facies begins to disappear. This major turn-over occurred mainly from the late Early Tithonian onto the Berriasian, a little later than the selected time slice of the Peri-Tethyan map. Therefore, the facies distribution on the Early Tithonian map is only a little modified compared with the Early Kimmeridgian: the carbonate platform environments have a larger extent and they dominate on the south Laurasia domain, from Transcaspian area to Iberia; they do the same all over the central Tethys and on the western Gondwanan shelves, except on the Arabian platform covered by evaporites. The disappearance of the radiolarites and the outburst of calcareous nannoplankton is regarded as a consequence of drastic change in marine water circulation due to a latitudinal flow through the central Tethys, the Maghrebian oceanic transfer zone and the central Atlantic onto the Caribbean domain (DE WEVER *et al.*, 1996; FOURCADE *et al.*, 1993, 1996). But, the accumulation of pelagic limestones during the Tithonian would be equally linked to palaeoecological, geochimical and physical modifications of the sea water, itself linked to the climatic evolution (BARTOLINI *et al.*, 1996; WEISSERT & MOHR, 1996). It is currently admitted that a warm and dry climate prevails during Late Tithonian on the large emerged areas of the intertropical zones (FOURCADE *et al.*, 1993). It appears that the climate should be slightly cooler along the southern border of the Laurasia and not so dry in the central Tethys platforms (BARTOLINI *et al.*, 1996; WEISSERT & MOHR, 1996). Opposite, it is clearly cooler on the Russian platform (RIBOULLEAU *et al.*, 1998).

Due to precise biostratigraphic correlations (HANTZPERGUE *et al.*, 1998a and b), it is worth noting that the stratigraphic distribution of Tithonian organic-rich deposits is different when comparing the UK - North Sea and the Russian platform. The probable variations of the water depth and water circulation, and the more or less important sandy input at several time are responsible for

the preservation, or non-preservation, of the organic matter. Therefore, the organic rich layers of the Kimmeridgian - Tithonian seam to be more linked to local or regional facies and palaeoenvironments, than to global oceanic anoxic events as in the Early Toarcian (BAUDIN & HERBIN, 1996; HANTZPERGUE *et al.*, 1998b).

III.- DEFINITION OF DOMAINS

III.1.- Russian platform: Volga - Ural - Donetz - Ukraine

The vast Russian epicontinental sea development undergoes strong modifications during the Early Tithonian (Early Volgian). The boreal seaway through the Sukhona - Vichegda, Mezen and Barents - Pechora basins is still opened; but, compared with its shape during the Callovian - Kimmeridgian interval, it is narrowed and divided into two branches by the Timan - Pechora uplift (VINOGRADOV, 1968). Its western part, Mezen - Upper Volga and Moscow basins, is underlined by clays and marls deposits. The intermittent marine seaway between the Pripyat - Dniepr - Donetz basins and the Baltic platform - Polish Trough no longer exists; as a consequence, the Voronezh, Ooka and Pripyat - Dniepr - Donetz basins form a wide gulf surrounded by the emerged and enlarged Fenno-Scandian shield, and the linked Ukrainian and Stavropol shields. The only direct remnant but reduced connection between the Tethyan and Boreal realms is through the Precaspian and Caucasus basins and the Turan plate, east bordered by the Ural highs - Kazakhstan plateau emerged areas.

The biostratigraphic study of type sections in the Middle Volga basin (HANTZPERGUE *et al.*, 1998a and b) makes henceforth possible to better correlate the Russian (MESEZHNIKOV, 1988b; MELEDINA, 1988, 1994) and the West European ammonites zonal schemes (HANTZPERGUE *et al.*, 1997). Similarly, Early - Middle Volgian strata which yield ammonites, radiolarians, foraminiferas and nannoplankton allow reliable datings in the Barents - Pechora basin (VISHNEVSKAYA *et al.*, 1999). But, the equivalence between the Volgian and/or Tithonian – "Portlandian" (*sensu anglico*) stages can only be suggested by affinities of the endemic Russian platform ammonites and the Boreal faunas. Thus, controversial stratigraphic datings remain, and the precise age of the facies recorded on the map would be still discussed. One of the major discrepancies which still remains is the age of the Volgian - Tithonian black shales of the Russian platform.

Marine deposition occurs in the majority of the sub-basins recognised over the Russian platform: the west Middle Volga, Volga - Kama, Ooka, Moscow and Upper Volga basins in the central part; the East Volga basin to the SE; the Pripyat, Dniepr - Donetz, Voronezh basins to the SW and the Sukhona - Vichegda, Mezen and Barents - Pechora basins to the NE. The relatively reduced sedimentation rate (thin series and condensed levels with phosphatic and pyritic nodules, glauconite and fossil accumulations), the presence of frequent unconformable surfaces (sometimes oxidised) and depositional gaps, witness evidence of shallow water and stability of the Russian platform. The sedimentation is mainly controlled by sea level variations; a more continuous deposition is observed toward the east Volga basin and the Caspian depression with indications of a more significant subsidence and deeper water (SAHAGIAN *et al.*, 1996). Everywhere, the basins subsidence is reduced without conspicuous faulting (STOVBA *et al.*, 1996; VAN WEES *et al.*, 1996; STEPHENSON *et al.*, 2000); it cannot be excluded that a later erosion should be too responsible of the reduced deposits or of their absence (ULMISHEK *et al.*, 1994).

The sedimentation is dominated by clays and silts; coarser sediment supply largely underline the uplifted Ural highs and the Stavropol - Ukrainian shields. In the gulf installed on the Voronezh - Ooka - Pripyat - Dniepr - Donetz areas, shallow marine limy and marly deposits with sands, grade into coastal plain to continental environments characterised by fresh-water ostracods and algae (Chara).

Palaeotemperatures data for the Early Volgian of the Russian platform are missing. However, in the Middle Volgian, the temperature fluctuates between 15 to 17°C, a slightly lower value than in the Kimmeridgian (RIBOULLEAU *et al.*, 1998). The continuous and regular but weak decrease of temperature since the Callovian is coeval with a global sea level rise which maximum is reached in Early Tithonian/Early Volgian. Such variations coincide with the southward increase of both Boreal faunal and sea-water influences while connections with the Tethyan areas and warm sea-water are more and more reduced; the isolation of the Russian platform epicontinental sea will be complete later, caused by the world wide Late Tithonian/Middle Volgian regression.

III.2.- Turan plate

The Tithonian is very difficult to separate, from the Kimmeridgian as well as from the Berriasian, because of very few significant biostratigraphic data and a great similarity of facies all along the Late Jurassic - Early Cretaceous series, deposited in shallow water marginal platforms and coastal plain environments (VOLOZH *et al.*, 1997). For example, Tithonian seas certainly covered the most part of the Turan platform and had connections with the Russian platform as indicated by "boreal" type brachiopod assemblage in NE Iran - Kopet Dagh (ADABI & AGER, 1997).

The sedimentation is dominated by varied limestones which extend north to the emerged Kara Bogaz high up to the Aral and NE Precaspian areas. Coral build-ups and evaporites are recorded in the Kopet Dagh range and Karakum depression; the terrigenous input (sandstones, silts and clays) is limited to the east, along the emerged Kazakhstan plateau. The absence of the Late Jurassic in several places, is certainly linked to the "Neo-Cimmerian" events which may be responsible for the uplifting of most of the Turan plate and subsequently the erosion of most of the Late Jurassic.

III.3.- Scythian platform - Crimea - Black sea - Caucasus - and Precaspian areas

The prominent rifting phase which characterises the Scythian platform and Crimea - Black Sea - Caucasus belt since the Middle Callovian - Kimmeridgian continues during the Early Tithonian (ROBINSON *et al.*, 1996; ROBINSON & KERUSOV, 1997; NIKISHIN *et al.*, 1998a and b, 2000). The sedimentation still shows a great diversity but this phase is marked by the most important evaporite event (Azov - Kuban, Terek basin and small depocentres in the Odessa Gulf). Carbonated deposits with build-ups still extend on platforms along the emerged Ukrainian and Stavropol highs (Fore-Caucasus, Azov - Kuban, Dzirula), separated from the baring areas by a narrow fringe of continental to coastal shallow marine clastic deposits; the thermal subsidence regime continues to be quiet and characterised by a decreasing rate (ERSHOV *et al.*, submitted). Active rifting phase and extension, with high rate of subsidence still plays in the north Black Sea belt, Great Caucasus trough and south Caspian areas which fill with thick marly siliceous sediments. Compression events continue in central and south Crimea (VOZNESENSKY *et al.*, 1998), as the manifestation of the Late Tithonian - general uplift and thrusting-folding.

Alternative palaeogeographic features would be used to illustrate these domains considering that a large back-arc deep-water basin with very thinned to local oceanic crust originated during Callovian - Late Jurassic (NIKISHIN *et al.*, 2000; BRUNET *et al.*, submitted). A crustal separation, followed by a sea-floor spreading would be possible; thus, a narrow oceanic crust might have been drawn in Great Caucasus trough, South Caspian and may be Kopet Dagh basins, between the emerged Central Iran - Alborz block and the South Turan rim plate.

III.4.- Teisseyre/Tornquist zone - Moesian platform

During the Tithonian (Early - Middle Volgian), the Polish - Ukrainian parts of the Teisseyre - Tornquist zone and the Moesian complex have a separate tectonic and sedimentary evolution. A regressive character is more and more marked in particular in the last phases of the Jurassic Polish basin development, while the Ukrainian Carpathians and Moesian areas still undergo a readily marine evolution.

In the Polish basin, continental areas progressively expand to such extent that the marine basin shrinks in a NW-SE trending; near the end of the Early Tithonian, the marine deposits remain only in the central and SE part of the Polish Trough (NIEMCZYCKA & BROCHWICZ-LEWINSKI, 1988; KUTEK, 1994; MATYJA & WIERBOWSKI, 1996; KUTEK & ZEISS, 1997; NIEMCZYCKA *et al.*, 1997; DADLEZ *et al.*, 1998; KUTEK, 2000). The communication with the Boreal sea is irregular, then interrupted because fluviatile-deltaic and coastal plain to shallow water sandy-clayey and sandy-marly deposits develop in Scania and the Danish basin. The communication with the Tethyan ocean is maintained, but only the southern part of the Polish domain belongs to the shelf of the Tethys. At the same time, the emerged Fenno-Scandian shield and the Bohemian massif enlarge (MALKOVSKY, 1987; ZIMMER & WESSELY, 1996), still being active sources of the clastic material which is shed into the reduced and narrow Polish trough. The corresponding deposits are shallow to deeper marine sandy or silty - clayey limestones, marls and shales in Pomerania and the central part of the Polish trough. No Tithonian deposits have been recognised on the Baltic platform (MAREK & GRIGELIS, 1998). Shallow platform carbonates and marls, more or less terrigenous, sediment in the south areas, along the border of the Ukrainian shield. The molluscs fauna is still a dominant feature, showing rare ammonites restricted to the central part of the Polish Trough; they allow to clearly recognise the Early Tithonian (Early Volgian). These ammonites which have boreal affinities indicate intermittent communications with the northern basins of Europe or the Russian platform only during the Early Tithonian (HANTZPERGUE *et al.*, 1998a), even if no intermediate marine sediments are preserved. A restricted seaway is supposed between the Pripyat - Dniepr - Donetz basins and the central Polish basin. The weak uplift events and the generalised regression which affects the whole northern Peri-Tethyan areas in the Late Tithonian should be responsible for the erosion of the pre-Cretaceous deposits; in all the discussed areas, the upper boundary of the Tithonian (or sometimes older successions) show an erosional character.

Alike in the Kimmeridgian, the subsidence axis of the henceforth shrank Polish basin is parallel to the limit of the East European platform (Tornquist - Teisseyre zone) and shifted from a NE position to the SW border which was previously the stable part. According to the different hypothesis of the tectonic control of the sedimentation, the Late Jurassic is either time of a rifting phase (KUTEK, 1994, 2000) with an extensional episode of increasing tectonic subsidence in its central and NW part (DADLEZ *et al.*, 1995; STEPHENSON *et al.*, submitted), or an episode of overall subsidence, not associated with high thickness gradients and with no evidences for synsedimentary activity (HACKENBERG & SWIDROWSKA, 1997; LAMARCHE *et al.*, 1998; LAMARCHE, 1999).

The palaeogeographic scheme of the Ukrainian Carpathians during Tithonian is very similar to that of the Oxfordian and Kimmeridgian (KUTEK, 1994; IZOTOVA & POPADYUK, 1996). The build up barrier is reduced and mainly bioclastic, the open sea deposits are more diversified with marls and silty - sandy limestones, and the back-reef system is composed of biodetrital and algal limestones. The scarceness or absence of coarse clastics and the direct present day boundary between the emerged Ukrainian shield and the shallow platform facies suggest that, either the shoreline was far away to the NE and the marginal marine facies have been later eroded, or that the relief of the emerged shield was very low.

In North Dobrogea, reverse-dextral slip, generated by a N-NE trending compression, occur along the Peceneaga - Kamena fault during the "Late Cimmerian" events, in the latest Jurassic - earliest Cretaceous (SANDULESCU *et al.*, 1995; ROBINSON *et al.*, 1996; HIPPOLYTE, 1996; BANKS & ROBINSON, 1997; BANKS, 1997; SEGHEDI, 2000). The Tithonian carbonate platform that previously fringed and probably partly covered the North and Central Dobrogea until the Early Cretaceous, is

progressively eroded because of the inversion tectonics. As a proof, the depositional environment of the Berriasian changes from a carbonate one in Tithonian, to a clastic dominated sedimentation in the deep shelves recognised in the present day western Black Sea near the Dobrogea (GEORGESCU, 1997). The assigned ages are based on phytoplanctonic associations, but the precision is not very good, and the distinction between Kimmeridgian and Tithonian still remains difficult. In the Predobrogea depression, the Late Jurassic keeps its shallow water calcareous, marly and dolomitic character alongside the emerged Scythian platform (MOROZ *et al.*, 1997), which may be underlined by fluviatile to deltaic deposits.

In South Dobrogea and in the Eastern Romanian plain which represents the biggest part of the Moesian platform, the sedimentation is carbonate dominated and seams to be more continuous but the precise dating is very difficult, mainly based on calcareous algae and charophytes assemblages (GRADINARU, 1993; AVRAM *et al.*; 1995). The depositional conditions are variable, running from shallow water platforms to coastal plain environments.

The Tithonian tectonic structures of the Bulgarian Moesian area is not strongly modified compared with the Kimmeridgian; the principal NW-SE fault set still delimits the Central Moesian basin from the West and East Moesian platform (SAPUNOV & TCHOUMATCHENCO, 1994). The former was an area of pelagic carbonated sedimentation with the predominant deposition of micritic limestones showing often similarities with the alpine "Maiolica" facies (SAPUNOV & TCHOUMATCHENCO, 1987, 1990; SAPUNOV *et al.*, 1985, 1988). The latter are the site of carbonate sedimentation, everywhere with build-ups on a ramp morphology; a north to south gradation of facies from shallow-marine through deeper ramp to basinal facies, a limited reworking of shelf material into the basin, intraformational slumpings, carbonate breccias derived from shallow-marine deposits and reworked into the basin, suggest a distally locally faulted and steepened ramp (HARBURY & COHEN, 1997). Similarly, syn-depositional normal faults controlled the subsidence of the SE margin of the Moesian platform, facing the Balkanides (SINCLAIR *et al.*, 1997). The eastern margin of the Moesian platform was apparently affected by crustal extension until the Early Tithonian, prior to the end Jurassic - earliest Cretaceous resumption of inversion movements, as indicated by the west to the east transition from shallow-water carbonate platform of the Bulgarian Varna area, to deeper water sediments in offshore Black Sea (DACHEV *et al.*, 1988). Compression structures of Late Jurassic age are noticeable too in the Strandzha range in Bulgaria (ROBINSON *et al.*, 1996). The Central Moesian basin spread on the southern rim of the platforms, prior to be incorporated into the Nis - Trojan flysch trough which constantly westward progresses during the Kimmeridgian and Tithonian (TCHOUMATCHENCO *et al.*, 1992; TCHOUMATCHENCO & SAPUNOV, 1994).

III.5.- Western Europe platform

The Early Tithonian of Western Europe corresponds as a whole to conditions of marine regression. However, the sea still occupies wide areas attested by the extension of ammonite faunas; but, because of such regressive tendencies, ammonites are affected by a strong endemism which characterises four areas (HANTZPERGUE *et al.*, 1997; GEYSSANT, 1997): the "Mediterranean province", on southern Iberia and Apulia, and the "Submediterranean province", on SE France and southern Germany, which are both parts of the "Tethyan domain"; the "French - German bioma", on Aquitaine and Paris basins and W Germany; the "Subboreal province" on U.K. and North France. Each of these provinces has its proper zonal scheme. Thus, if in the literature biostratigraphic record is imprecise, for instance prior to the henceforth published correlated biostratigraphic and chronostratigraphic schemes (HANTZPERGUE *et al.*, 1998a and b), or if authors indicate only "Tithonian", "Volgian", "Portlandian", "Kimmeridgian", without referring to an up-to date stage definition (especially "Kimmeridgian *sensu gallico* or *sensu anglico*), the collect of reliable data is very difficult and somewhat confusing.

The Central Atlantic spreading system is very active and connected with the Iberia - Africa transfer zone (VERA, 2000); but, the separation along the W Iberian margin has not yet begun. The North Sea is still undergoing rifting activity with the onset of a major extension in its central grabens (RATTEY & HAYWARD, 1993; PARTINGTON *et al.*, 1993a and b; UNDERHILL & PARTINGTON, 1993), showing the associated tectonic, eustatic and sedimentary events which illustrate the last episodes of the "North Sea Cycle" (JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998). The rifting story of the North Sea is going to its end. The extension is transferred onto the Atlantic margins.

The beginning of the Late Jurassic regression is everywhere evident and the marine communications between the Boreal and the Tethyan realms become restricted if not totally interrupted (Hanover basin, Polish Trough) prior to the Late Tithonian general emersion of the W Europe. The whole area undergoes the effects of a combined general uplifting and sea-level fall which in many places reduce the accommodation space, as evidenced by the development of continental or brackish coastal plain environments on the borders of several widening emerged areas (Iberian, Armorican, Central, Rhenish, Bohemian massifs).

In the Northern Europe (U.K. and North Sea), the sedimentation is dominated by fine clastics distal platform deposits, several time interbedded with organic rich shales and coarse clastic layers. Shallow marine platform or ramp carbonates with frequent build-ups, sometimes grading to coastal environments with fluctuating salinity, develop on the whole Southern Europe; they northward extend all around Iberia, in the north Aquitanian basin, the southern Paris basin, the southern rim of the Bohemian massif and the central Polish Trough. Local coastal marine to lagoonal environments, on the fringe of dry lands, are evidenced by dinosaurs foot-prints (northern border of the Aquitanian basin, Provence platform and North Spain).

The facies succession of the Lusitanian - Atlantic margin of Iberia grades simply from continental and brackish terrigenous sediments on the east, to shallow marine platform carbonates and deeper marls to the west. Because of a poor fossil record, these deposits, which

overlie marine sediments dated Late Kimmeridgian (*sensu gallico*), have been often referred to "Portlandian" (*sensu gallico*). On the offshore Iberia and Galicia, platform carbonates underline the present day shore-line onto the Algarve. In spite of several active normal faults, the subsidence appears to be mostly regional, in a "pre-rift" context, than controlled by a tectonic activity (WILSON, 1988). The main rifting events and the break-up of the Iberian plate will occur not before the Late Tithonian - Berriasian with a northward propagation along the W Iberia margin (WILSON *et al.*, 1991) supposed to be related to a "super-plume" (WILSON, 1997).

The palaeogeography of NE Iberia illustrates an episode of the rifting cycle which spans from Latest Oxfordian to Late Hauterivian (CANÉROT, 1989, 1991; SALAS *et al.*, 2000). The "Soria seaway" is interrupted by syn-rift continental deposits in the Cameros basin which is part of a set of extensional sub-basins structured by a normal fault system. Rift-induced subsidence and tectonic uplift of the edges have started soon in the Early Oxfordian in the Maestrat basin; thus, Early Tithonian platform carbonates and shallow marine terrigenous or carbonated deposits make a SE to NW transition with the continental facies. The relative sea-level variations show a constant fall (BÁDENAS & AURELL, 1997; AURELL *et al.*, 1998, 2000).

In the Pyrenees (CANÉROT, 1989; 1991; VERGÉS & GARCÍA-SENZ, 2000) and the southern border of the Aquitaine basin (LE VOT *et al.*, 1996; JAMES, 1998) the Tithonian deposits settle on a very shallow and stable ramp periodically emerged. Carbonated facies prevail with frequent regional hiatuses, dissolution marks, interbedded breccias and stromatolite crusts. From the latest Tithonian to Barremian, the eastern part of the shelf is uplifted and subject to erosion as evidenced by local hiatuses. In the western part of the basin, sedimentation is more continuous, showing a west gradually deepening. The northern border of the Aquitaine basin is underlined by shallow carbonated shaly facies; marine communications with the Paris basin through the so-called "Poitou trough" seams to be interrupted, prior to the general Late Tithonian ("Late Portlandian" *sensu Gallico*) emersion.

The Dauphinois basin, eastward widely opened on the Valais trough and the Alpine sea, still undergoes a marly to shaly deep-hemipelagic sedimentation; it is surrounded by the Jura, Cévennes - Languedoc and Corbières - Provence carbonated shelves, with several limited coral-reef build-ups. It is admitted that marine communications would exist with the Aquitanian basin, and probably with the wide platform carbonate facies which extend between the Ebro massif and the present day Pyrenées chain.

The Paris basin is occupied by shallow to deeper marine aggrading platforms characterised by "Early Portlandian" (*sensu gallico*) marly limestones and marls interbedded with micritic limestones; shallow to coastal marine terrigenous facies exist in the Boulonnais, the only witness to the near emerged London - Brabant massif. The tectonic control of the sedimentation still exist but, alike the Kimmeridgian, the Early Tithonian would be a moment when the subsidence speed is maximum and homogeneous all over the basin (GUILLOCHEAU *et al.*, 1999; ROBIN *et al.*, 2000). The Jura platform would be certainly connected with the carbonated platform of the Paris basin; but, a latest Jurassic - Early Cretaceous erosional phase has removed the "Portlandian" deposits everywhere on the periphery. Intraplate deformations and a thermal event would be responsible for the general uplifting of the Paris Basin during the Late Tithonian - Early Berriasian interval (GUILLOCHEAU *et al.*, 1999; ROBIN *et al.*, 2000). In addition to the erosional phase, the Jurassic / Cretaceous boundary is underlined by an angular discordance, the "Neo-cimmerian unconformity", within the so called "Purbeckian facies".

The Swiss Jura platform is prolonged to the east by shallow water platforms onto Schwabia and Franconia. These are Late Jurassic stable palaeogeographic elements with marine environments until the earliest Tithonian. But, the northern termination of the so-called "Hessische strait" will soon become closed by brackish coastal plain, lagoonal and continental terrigenous deposits which fill the Hanover - Lower Saxony basin (KÖLBEL, 1968; KLASSEN, 1984; GRAMANN *et al.*, 1997) onto the German North Sea grabens (KOCKEL, 1995). Alike during the Kimmeridgian, South Germany is characterised by a micritic or coarse grained bioclastic limy sedimentation; sponge bioherms are well developed on the Franconian platform, while they are less homogeneous and more irregular in Schwabia (MEYER & SCHMIDT-KALER, 1989; BRACHERT, 1992). At the end of the sponge-reef development, additional corals, provide evidence of the shallowing of the South Germany shelf sea.

As a continuation of the Kimmeridgian, the Tithonian palaeogeographic evolution of the northern NW Europe is controlled by the relative sea-level variations of the "North Sea cycle" (STEEL, 1993; PARTINGTON *et al.*, 1993 a and b; UNDERHILL & PARTINGTON, 1993; JACQUIN & GRACIANSKY, 1998; JACQUIN *et al.*, 1998) and by the last episodes of the failed North Sea rift. The map locates at time just before the relative quiescence of the northern North Sea (RATTEY & HAYWARD, 1993), within the early part of the "Late Kimmeridgian" (*sensu anglico*). In this case, most of the deposits concerned are the uppermost part of the "Lower Kimmeridge Clays", the clayey facies yielding laminated bituminous shales and sandy layers as evidenced in the type areas of Dorset (CALLOMON & COPE, 1995; WATERHOUSE, 1995). Thus, the Early Tithonian palaeogeography of the U.K. and the North Sea is very similar to that of the Early Kimmeridgian (Late Kimmeridgian *sensu anglico*, Early Volgian; COPE & RAWSON, 1992). The "Kimmeridge clays" or correlated-equivalent formations are reported from the hinterland west Weald basin, the London platform, the east Middland shelf, the Yorkshire - Cleveland basin and numerous offshore areas of the North Sea (BROWN, 1990). Keeping in mind the relative biostratigraphic impreciseness and possible confusing chronostratigraphic understanding of the offshore records, the most informative data refer to the Western Approaches basin and the English Channel (HAMBLIN *et al.*, 1992), the Cardigan bay and Bristol channel (TAPPIN *et al.*, 1994), the Moray Firth (ANDREWS *et al.*, 1990; CASEY *et al.*, 1993; STEPHEN & DAVIES, 1998), the southern and central North Sea (HERNGREEN & WONG, 1989; CAMERON *et al.*, 1992;

CLARCK *et al.*, 1993; DONOVAN *et al.*, 1993; PRICE *et al.*, 1993; STEEL, 1993; WAKEFIELD *et al.*, 1993; VAN ADRICHEM-BOOGAERT & KOUWE, 1997; GATLIFF *et al.*, 1994; KOCKEL, 1995), the Norwegian North Sea - Ergesund - Danish basin (JOHANNESSEN & ANDSBJERG, 1993; STEWART, 1993) and Viking graben (GARLAND, 1993; STEEL, 1993).

The sedimentation is currently fine grained in the S areas and in the central North Sea with clayey-mudstones and calcareous-mudstones, interbedded with silts, silty-mudstones and siltstones-clays deposited in marine low-energy environments; bituminous layers and thin limestones beds are intercalated into the whole formation. Medium to coarse grained facies occur frequently in the north, overall near the emerged areas; these are deposited in submarine fan-like complexes or shallow marine to coastal marine terrigenous platforms. In the SE part of the North Sea, continental to fluviatile and coastal plain coarse grained terrigenous facies are particularly well developed in small grabens between the emerged London Brabant massif and the enlarged Ring Kobing Fyn High; they close the former marine communications with the central Tethys domain. Pebbly detritus layers are frequently interbedded in finer deposits of the Central Graben (HUMPHREYS *et al.*, 1991), in southern Britain and northern France (GARDEN, 1991); they indicate a Triassic, Permian and Carboniferous provenance of the near uplifted and emerged blocks which undergo a strong erosion, reinforced by the general Late Jurassic regression.

III.6.- Maghreb (Morocco - Algeria - Tunisia) - Saharan areas

The main palaeogeographic features of the Kimmeridgian are emphasised during the Early Tithonian onwards. The tilted blocks structured margin, with northward vergence (CHOTIN *et al.*, 2000), still drives to the increasing of deepening and subsidence of the Tell (BENEST *et al.*, 1993; ATROPS & BENEST, 1994) and Rif areas (VIALLY *et al.*, 1994; BRACENE *et al.*, submitted; ELLOUZ *et al.*, submitted) underlined by a widespread deposition of nodular limestones, marls and marly limestones (BENEST & GHALI, 1985; BENEST, 1990; ATROPS & BENEST, 1993, 1994; BENEST *et al.*, 1993).

At the same time, regressive deltaic and coastal plain to coastal marine environments extend into the Atlasic (High and Middle Atlas) and Saharan domains (Lower and North Sahara basins, Oued Mya and Dahar basins), as far as the Tataouine and Ghadames basins. The lithology and the fossil record suggest a persistent humid climate (LEFRANC & GUIRAUD, 1990; BUSSON & CORNÉE, 1991). Terrigenous input (sands and silts) is a constant and outstanding feature of the various calcareous, dolomitic and marly shallow marine facies that develop in marginal basins (Essaouira, Middle Atlas) and in the Saharan domain, frequently with evaporitic intercalations. Opposite, coastal to shallow marine platform carbonates, sometimes with coral build-ups, and marly limestones are dominant from the north Middle Atlas to the Chotts area and Gulf of Gabès (BELHAJ, 1996), through the Lower Sahara basin.

From the Late Kimmeridgian to Tithonian - Early Berriasian, a carbonate rimmed shelf and ramp develop in the Rif areas (FAVRE *et al.*, 1991), characterised by fine grained sedimentation in very shallow water; sedimentation is still guided by extension along a NE - E-NE and SW - W-SW trending, linked to the general divergence between Africa and Iberia (CATTANEO, 1991; ZIZI, 1996). In the Middle Atlas, the northern fringe of the deltaic series is drowned by the last Jurassic sea level-rise, contributing to maintain the Late Jurassic carbonate platform which surrounds the Maghreb, without interruption from the Essaouira basin up to the Gulf of Gabes, through the Middle Atlas, Oran High Plains and the Tlemcen - Preatlasic domains.

Deeper and pelagic facies (nodular and siliceous limestones), sometimes silty and clayey (marly limestones and marls), extend along a narrow fringe west and north to the Atlantic Morocco and Middle Atlas; they widely develop in Tunisia, with pelagic fauna (radiolarians) but without or very feeble and fine clastics (North-South Axis and Tunisian Dorsale; PEYBERNÈS *et al.*, 1990; SOUSSI *et al.*, 1998, 1999 and 2000). Extensional N-S activity plays during the Tithonian, related to the constant Africa-Eurasia divergence (BARRIER *et al.*, 1993; BOUAZIZ *et al.*, 1996, 1998, 1999). The moderate increasing subsidence rate revealed in South Tunisia should be due to the beginning of a rifting phase which will strongly act in Early Cretaceous (PATRIAT *et al.*, submitted).

In the Essaouira basin, no reliable data allow to clearly distinguish the Tithonian from the Kimmeridgian; the upper part of a thick marly series with dolomitic and evaporitic intercalations, is often referred to the Tithonian *s.l.*. (FAVRE & STAMPFLI, 1992; BROUGHTON & TRÉPANIER, 1993; MEDINA, 1994; MORABET *et al.*, 1998). Outstanding facies variations show a more marked sandy input on the north, gypsiferous marls in the centre and dolomitic to limy layers to the south. Platform carbonates and deeper carbonates (sometimes with calpionellids) are found in offshore boreholes; then, the recorded facies should upgrade until the Late Tithonian - Early Cretaceous. The subsidence rate is still high within the distensive framework of the Moroccan Atlantic margin (MEDINA, 1995; LABBASSI *et al.*, 2000; CHOTIN *et al.*, 2000).

Therefore, the coeval deepening of the Maghreb Tethyan margin and its Atlantic façade, run counter the eustatic regressive trends that generally characterise the uppermost Jurassic time: the Tethys and Atlantic spreadings lock the Atlasic spreading.

III.7.- Egypt - Sudan - Libya

The Tithonian stage is often very difficult to identify because reliable biostratigraphic data are missing in many places; many times, the Tithonian cannot be distinguished from the Kimmeridgian in the monotonous platform carbonate successions which develop along the Tethys margin (GUIRAUD *et al.*, 2000). Tectonic activities increase, as evidenced by magmatic intrusions in Sinai, Sirt basin, South Egypt and North Sudan (WILSON & GUIRAUD, 1998). A strong tectonic event, the "Neo-Cimmerian", occur near the Jurassic / Cretaceous transition, underlined by unconformities and gaps in the sedimentation; moreover, most of the Jurassic series should be eroded in conjunction with the tectonic inversion that occurred at that time and later in the

Cretaceous. Subsequently, it is often difficult to clear up what can be referred to Tithonian (and subsequently to its lower part), within the successions described as comprehensive "Kimmeridgian - Portlandian" or "Kimmeridgian - Tithonian".

The Tithonian is absent in south and central Sinai (JENKINS, 1990) but it is recorded in offshore Sinai and west Sinai - north Gulf of Suez (KERDANY & CHERIF, 1990; DARWISH, 1992); this subsident area is a continuation of the Late Jurassic basin extending over the northern part of the Sinai and northern Eastern Desert, up to the Nile Delta and the northern Western Desert. The deep seated normal faults, parallel to the present coast, continue to actively control the sedimentation which grades, following a S-N succession of belts, from continental - fluviatile to coastal plain, and marine shallow inner to deeper platform environments. In the latter, palynological assemblages, recorded in boreholes of NW Egypt, emphasise that no major hiatuses are recognised throughout the succession at the Jurassic/Cretaceous transition (MAHMOUD *et al.*, 1999), as inferred from previous investigations in this areas (ABDEL AAL *et al.*, 1990; KEELEY *et al.*, 1990). This suggests that the several syn-depositional tectonic elements recognised in northern Egypt, have separate sedimentary records, regionally and locally controlled by a W-E normal faulting brought about by SW-NE stretching (KEELEY & WALLIS, 1991; GUIRAUD & BELLION, 1996; KEELEY & MASSOUD, 1998; GUIRAUD, 1998; GUIRAUD & BOSWORTH, 1999).

In the Sirt basin, the basal clastics of the "Nubian sandstones" sequence, lying on the basement (EL-HAWAT, 1992; EL-HAWAT *et al.*, 1996; WENNEKERS *et al.*, 1996) are inferred to involve the comprehensive "Kimmeridgian - Tithonian" interval. The stratigraphic sequence exhibits several steps of sedimentation in a typical rift basin with continental-fluviatile to coastal plain and coastal marine environments; the sedimentation is controlled by NW-SE normal faulting (ANKETELL, 1996). The correlative platform carbonates and deeper marly facies are recorded in offshore Sirt and Benghazi basins (SMITH & KARKI, 1996), which are the continuation of the offshore Cyrenaica formations (DURONIO *et al.*, 1996). The sedimentation is controlled by normal faults parallel to the present coast line; the offshore Sirt basin half grabens, possibly of Late Jurassic, and the Benghazi basin, left lateral pull apart (SMITH & KARKI, 1996), suggest that tectonics is associated with an episode of thermal bulging and stretching along, with the deposition of syn-rift sediments.

Alike for the Early to Middle Jurassic, the Tithonian interval may be registered in the continental-fluviatile "Nubian Sandstones" of the Dakhla, Al-Kufra and Erdis basins; these ones are assumed to be linked, leaving reduced uplifted areas on Mid Cyrenaica and Uwainat high, Tibesti-Gargaf and Arabo-Nubian shields.

III.8.- Levant (Israel - Lebanon - Syria - Jordan)

The Tithonian palaeogeography of the Levant is very speculative because this interval is badly dated and a general uplift generates the infra-Cretaceous unconformity which is marked by wide denudation and the Tayasir volcanics in Samaria and Galilee, Lebanon and Golan (LAWS & WILSON, 1997; HIRSCH *et al.*, 1998). A marine pelitic turbidite-like facies develops in the Gevar'Am trough alongside the present Levant coast; it is probably related to active faulting along the passive margin, initiating a separation between an east Levant platform (Arabo-Nubian) and a western shallow marine platform (East Mediterranean). Nearly all Callovian to Upper Jurassic formations are missing, both by subaerial truncation (central and south Negev) and channelling of the turbidites facies (present day coastal plain and Judea).

In Lebanon thick limestones record the major Late Jurassic eustatic rise which probably culminates in the Early Tithonian interval (WALLEY, 2000). These mid shelf to near shore deposits are overlain by similar facies which represent deposition in a very shallow and high energy water conditions at the end of the Jurassic. The total sedimentary pile illustrates a low order transgressive - regressive cycle which may relate to a temporary waning in Late Jurassic tectonism, underlined by the "Bahnes" magmatism episode. Alike in Israel, the Jurassic - Cretaceous boundary is again marked by major tectonic phase; a general uplift and block faulting drive to erosion of a more or less important part of the Middle to Late Jurassic sediments.

While the western margin of the Mount Lebanon - Syrian platform continue to subside until the end of Kimmeridgian (shallow marine deposits), the latest Jurassic regressive phase begins and the entire Syrian platform becomes progressively emerged. The Tithonian is missing in Syria and probably in all the neighbouring areas. A large emerged promontory extend in a N-E direction on the Arabo-Nubian shield, joining the Mardin, Hamad - Rutbah and South Negev highs.

III.9.- Central Arabian platform and Gulf area - Iraqi platform - Oman - Zagros basin

The tectonically active period at the end of the Jurassic (break-up at South Tethyan margin and drifts of India from Gondwana) has not yet marked effects upon the Arabian plate during Early Tithonian; the palaeogeography is not strongly modified in comparison with the Kimmeridgian. The tectonic impact remains weak in Saudi Arabia and surroundings while the platform margin in Oman and Yemen is a little more concerned: extensional phase with flexure which resulted in drowning of the margin; doming which affected the interior zone of the platform. In the latter, the biostratigraphic and sedimentary data (calpionellids and foraminifera; siliceous pelagic and "maioloica-like" facies) indicate that the detrital influx and marine openings take place in Late Tithonian - Berriasian; subsequently they are not reported on the Early Tithonian map.

In central platform and gulf area, the Tithonian sedimentation is mainly controlled by eustatism, locally reinforced by intraplate tectonic marked by S-SW - N-NE synsedimentary normal faults, orientated in a parallel direction to the passive Arabian plate margin (LE NINDRE *et al.*, submitted). The depot centre is still clearly located

in Central Arabia but a southward thickening of deposits is registered since Early Kimmeridgian to Tithonian. The evaporitic formations (Le Nindre *et al.*, 1987), as well as their equivalent in Iraqi - Iranian areas, remain undated because the biostratigraphic data are totally missing; in spite of still confusing correlations, these extensive formations are considered as Tithonian (Grabowsky & Norton, 1995; Carman, 1996; Meyer *et al.*, 1996; Hughes, 1996; Youssif & Nouman, 1997; Alsharhan & Nairn, 1997; Al Husseini, 1997). Witness to the Late Jurassic regressive trends, they are the outstanding palaeogeographic feature of the Arabian plate (Murris, 1980), typical of sabkhahs deposits, from intertidal and proximal to littoral palaeoenvironments under evaporitic conditions. Little marine invasions over the platform are registered in occasional limestone and shales which form one of the main producing reservoir, interbedded into salt and anhydrite facies. Refering to a sequence stratigraphy approach (Le Nindre *et al.*, 1990b), they represent four major sequences (Hughes, 1996; Al-Husseini, 1997; Youssif & Nouman, 1997), subdivided into higher frequency sedimentary cycles (Meyer *et al.*, 1996).

In Interior Oman and Oman Mountains, the shallow marine carbonates, and their equivalent in Zagros fold-belt, are supposed to span until the Early Tithonian according to the age given by foraminiferas (Le Métour *et al.*, 1995). The beginning of the thermal doming activity linked with the rifting of the Indian Ocean, certainly induces the uplift of Dhofar High and eastward emersions, as well as SW-NE normal faulting in Oman; moreover, Tithonian deposits may have been partially eroded in several places.

III.10.- Ethiopia - Somalia - Gulf of Aden - Yemen

The Tithonian sediments are very difficult to distinguish from that of Kimmeridgian. Due to the Late Tithonian regression, a considerable part of the Late Jurassic must have been removed by Early Cretaceous erosion; for example on the Erigavo high and partly in the Ahl Mado and Ahl Meskat basins, Jurassic pebbles are reworked within the commonly distributed basal conglomerate deposited during the Cretaceous transgression (Luger *et al.*, 1990, 1994b). Generally, despite of few biostratigraphic data (Mette, 1993; Simmons & Al-Thour, 1994), Early Tithonian is defined within the sedimentary piles between the dated Kimmeridgian and the Late Tithonian which yields Calpionellids associations (Luger *et al.*, 1994b). The Central Ethiopia and the North and West Somali basins undergo a reactivated subsidence in rifting conditions along troughs associated with the extension resulting from the separation of the Indian plate from the African. A westward thickness increase indicates that the structural differentiation observed in Callovian - Kimmeridgian still plays in Tithonian. For example, the westerly areas of the Ahl Mado basin display higher subsidence rates of carbonaceous shallow marine nearshore environments than the easterly, which remain in more marly pelagic conditions (Bosellini, 1989; Bosellini *et al.*, 1997; Luger *et al.*, 1990, 1994b).The scarceness or absence of coarse siliciclastic input in marine sediments of NW Somalia and Central Ethiopia suggest the lack of high and significant areas of erosion. However, in the NW and central part of the Plateau of Ethiopia (Upper Abbay River basin), the Tithonian should be represented in the more siliciclastic limestones and the shaly sandstone units (Getaneh, 1991) inferred to be deposited in supratidal environments, grading into a meandering river system.

In North Yemen, in the Sana'a area, formations are dated as Late Kimmeridgian - Tithonian by large foraminiferas (Al-Thour, 1997); the predominantly carbonated deposits with intercalated sandy-silty dolomitic and marly shales are correlated with shaly evaporitic to continental and intertidal to fluctuating salinities deposits in South Yemen (Beydoun, 1989, 1997; Beydoun *et al.*, 1996), which may extend onto the Socotra area, eastern Gulf of Aden (Birse *et al.*, 1997). No direct marine communication seams to be possible with central Arabian platform.

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12.- EARLY HAUTERIVIAN (123 - 121 Ma)

L. G. BULOT (COORDINATOR)[1]

with scientific contributions of
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I.- MAIN FEATURES

Time slice definition and biochronology

The Hauterivian stage was originally introduced by RENEVIER (1874), based on lithostratigraphic units occurring in the type area of Hauterive (Neuchâtel, Switzerland). Since then, the concept and limits of the stage have evolved, and as most Jurassic and Cretaceous stages, the Hauterivian and its subdivisions are defined on the basis of ammonite biochronology.

As for other early Early Cretaceous stages, a distinctive provincialism is recognisable for the Hauterivian, resulting in two different zonation schemes: a Tethyan and a Boreal standard zonation. The Hauterivian stage is, however, marked by a pronounced exchange of marine faunas and floras between Tethyan and Boreal Realms, while the Berriasian, Valanginian and Barremian stages are characterised by more endemic taxa. Thus, within Europe, some markers are present in both realms, allowing inter-regional correlation.

Moreover, there is also a consensus that the Hauterivian stage should be defined in the Mediterranean region of the Tethyan Realm and thus, the key section at La Charce (Drôme, France) is recommended as the boundary stratotype (GSGP). As defined in recent literature, the stage ranges from the base of the Radiatus ammonite Zone to the base of the Hugii ammonite Zone (BULOT, 1996; MUTTERLOSE, 1996; RAWSON, 1996). With respect to the present map, the authors suggested that the early Hauterivian should be selected as a key period because it correspond to the peak extension of a marine transgression that started in the Valanginian (see regional synthesis below).

As considered herein, the Early Hauterivian substages includes the Radiatus, Loryi and Nodosoplicatum ammonite Zones and ends at the base of the Sayni ammonite Zone. It should be noted that the early - Late Hauterivian boundary is still under discussion as the LO of the nannofossil *Cruciellipsis cuvillieri* was proposed as an alternative to the base of the Sayni Zone (MUTTERLOSE, 1996). In any case, the reference sections recommended are located in the deep-water facies of the Vocontian basin (SE France) (BULOT *et al.*, 1993). Therefore, the ammonite scale developed in SE France stands as a standard for the Mediterranean region of the Tethyan Realm (HOEDEMAEKER & BULOT, 1990; HOEDEMAEKER *et al.*, 1993).

The North European standard ammonite scale proposed by RAWSON (1983) was established after the reference sections at Speeton (Yorkshire, UK), Moorberg - Sarstedt (North Germany) and Pollhagen (North Germany) (RAWSON, 1971, MUTTERLOSE, 1984). Correlations between the North European and Mediterranean ammonite scales are supplemented by the occurrence of Tethyan migrants in the subboreal ammonite successions of North Germany and Great Britain (BULOT & THIEULOY, 1995; RAWSON, 1995, BULOT *et al.*, in prep.).

Direct calibration of various micro- and nannofossil groups on the Hauterivian ammonite scales have been carried out over the last decade (BERGEN, 1994; MAGNEZ-JANIN, 1995; MUTTERLOSE, 1996; LEEREFELD, 1997; BULOT *et al.*, in prep.). Significant events for the definition of the early Hauterivian can be summarised as follows:

- the LO's of the calcareous nannofossils *Percivalia bullata* and *Eiffellithus windii* are close to the base of the Radiatus Zone (Tethyan Realm only);
- the FO of the calcareous nannofossil *Litraphidites bollii* falls within the middle part of the early Hauterivian Loryi Zone (Tethyan Realm only);
- the range of the calcareous nannofossil *Eprolithus antiquus* approximate the duration of the early Hauterivian (boreal Realm only);

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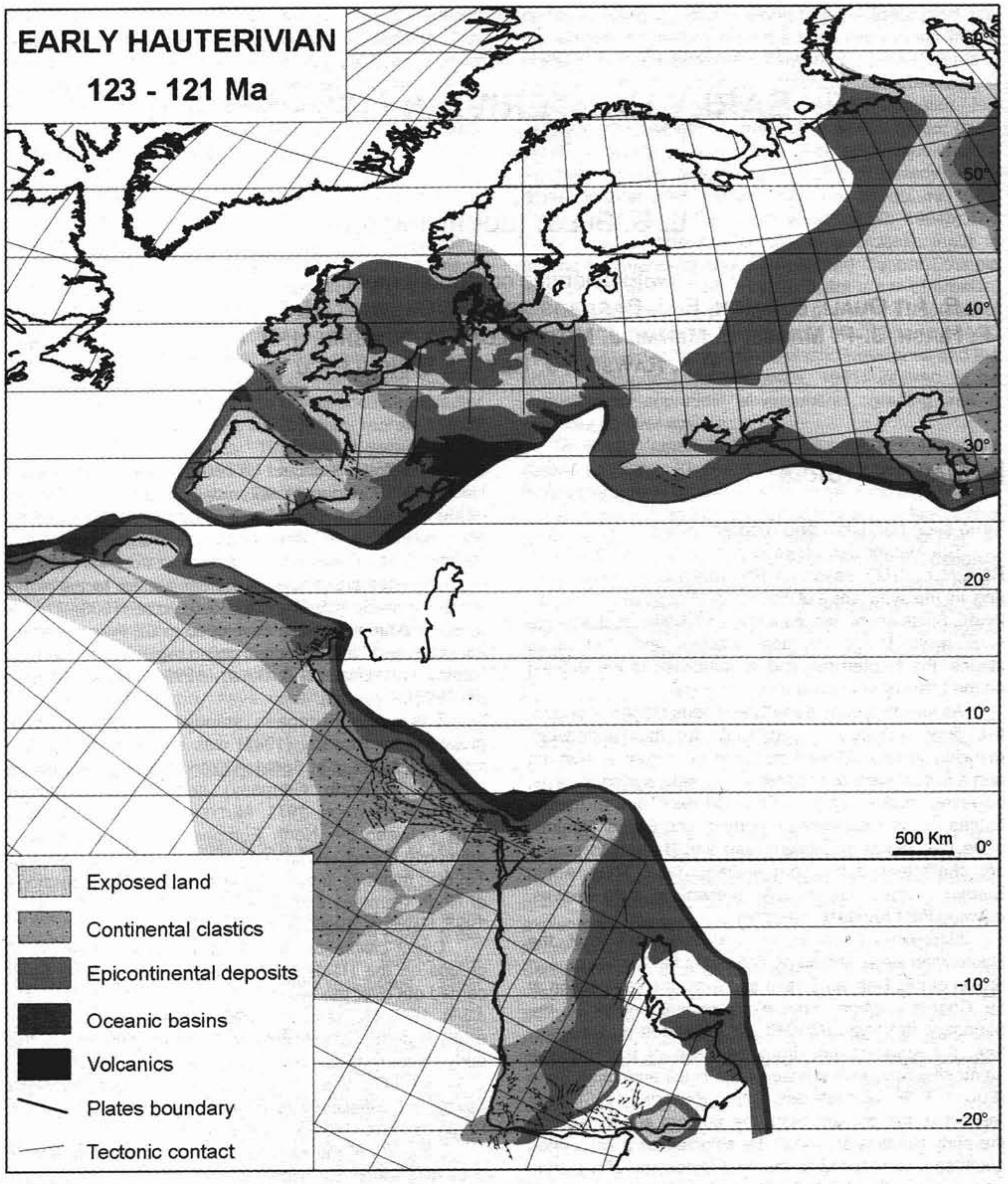

Fig. 12.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Hauterivian.

- the FO of the dynoflagellate *Muderongia staurota* occurs in the upper part of the Radiatus Zone (Mediterranean area) and at the top of the Amblygonium Zone (North Germany).

In shelf environments, it should be noted that the ostracod *Protocythere triplicata*, formerly considered as a marker of the Early Hauterivian (Oertli, 1989), has now been proved to occur already in the middle to upper part of the Late Valanginian (Bulot, 1995; Mutterlose, 1996; Mojon, unpublished data). Similarly, the irregular echinoids *Toxaster amplus* and *Toxaster retusus*, often mentioned in the literature as Hauterivian index fossils, should be reconsidered in the light of new taxonomic work (David, 1979; Clavel, 1989). Even if carbonate platform settings show a limited extension during the Early Cretaceous (see below) large foraminifera, dasycladales and rudist provide useful markers (Masse, 1993 and 1994). It should be noted that most Early Hauterivian

dasycladales and foraminifera are suspected to have already appeared in the Late Valanginian (BULOT *et al.*, 1997).

Non-marine Early Hauterivian can be identified by means of lacustrine ostracods (Assemblage 7 of *Cypridea*; ANDERSON, 1973) and charophytes oogons. By contrast, palynological characterisation of the non-marine Hauterivian is still problematic despite the tremendous work carried out on the upper part of the Wealden and equivalent formations of Northern Europe. In our present stage of knowledge, neither miospores nor pollens allow the establishment of a discrimination between the Valanginian, Hauterivian and Barremian stage (see discussion and references *in* BATTEN, 1996).

Absolute ages for the Hauterivian vary very much according to authors (see discussion *in* GRADSTEIN *et al.*, 1994). Nevertheless, most recently HENNIG *et al.* (1999) have shown that the Valanginian - Hauterivian boundary fall within the normally magnetised reversal CM10. This implies revision of the correlation proposed by GRADSTEIN *et al.* (1994, 1995).

II.- STRUCTURAL SETTING AND KINEMATICS

II.1.- Plates and blocks accounted for the Early Hauterivian

The opening of the southern Atlantic marks the Late Jurassic to Early Cretaceous period. As a consequence, major palaeogeographic changes affected the Tethys, while Africa and South America were moving northwards. The Tethys recorded compressive tectonic events between North and South America, and between the Arab-African block and Eurasia. This event, known as the Neo-Cimmerian or Nevadan phase (140-125 Ma), is characterised by the onset of Cretaceous flysch that indicates the beginning of the orogenic phase in the Tethyan areas. In the meanwhile, the East Gondwana plate broke up into the Indian and Australian plates. The transit plate is also assumed to have split into two parts during the same time interval (DERCOURT *et al.*, 1993). As a whole, the history of the Early Cretaceous Tethys is marked by periods of high geodynamic activities, sometimes very short, alternating with relatively longer, quiet phases.

II.2.- Palaeoposition of plate and blocks

The present map was obtained by applying constant sea floor spreading to the palinspastic reconstruction of the Late Tithonian previously established by FOURCADE *et al.* (1991, 1993a and b). The Hauterivian map obtained fits in well with the reconstruction of ZIEGLER *et al.* (1987). Nevertheless, it should be noted that this latter map was originally dated as Valanginian, but the reconstruction was obtained from magnetic anomaly M10, that is now assumed to fall within the Early Hauterivian (HENNIG *et al.*, 1999).

FOURCADE *et al.* (1991, 1993a and b) have already presented a methodology for dynamics and reconstruction of palaeolatitudes. An independent source of data with an extensive reference list is also available from BARRON (1987). It should also be noted that the reconstruction of palaeolatitudes by ZIEGLER *et al.* (1987) was accomplished by the combination of two different sets of data, e.g., palaeolatitudinally sensitive sediments and palaeomagentic pole data. Techniques, references and details are to be found in those papers.

II.3.- Accuracy

Despite considerable improvement in the correlation of the M-sequence polarity chrons to palaeontological events and zones, the absolute age control on the Late Jurassic and Early Cretaceous remains poor. Consequently, CHANNEL *et al.* (1995a) consider that the available absolute age control is insufficient to justify abandoning the constant oceanic spreading rate assumption for that part of the geological time. Moreover, comparison of block models from Japanese, Hawaiian, Phoenix and Keathley lineations suggests that the Hawaiian model is the closest to a constant spreading record. It should also be noted that the Hawaiian model presented by CHANNEL *et al.* (1995a) differs slightly from the published data of LARSON & HILDE (1975). Moreover, recent integrated bio-, chemio- and magnetostratigraphy in SE France and Italy has shown that the Valanginian/Hauterivian boundary falls within the normally magnetised zone CM10 (HENNIG *et al.*, 1999). The same boundary was formerly placed by CHANNEL *et al.* (1995b) and MUTTERLOSE (1996) into CM11 and dated 131.5 Ma. This age is most likely to be re-evaluated. Besides, the base of CM9 still stands as a good proxy for the Early/Late Hauterivian boundary (CHANNEL *et al.*, 1995a and b).

II.4.- General comments

The tectonic features presented on the map have greatly benefited from the unpublished reconstruction by Joseph CANÉROT (Portugal, Spain, Morocco), Francis HIRSH (Israel) and René GUIRAUD (Africa). The structural evolution of Africa (and the Middle East) during the Early Cretaceous controlled the nature of sedimentation, both of terrestrial and epicontinental marine facies (GUIRAUD & MAURIN, 1992; GUIRAUD, 1998). The marine strata developed primarily from the peripheral basins. Non-marine sediments are found, in addition to the peripheral basins, in many inland areas (see overview by MATEER *et al.*, 1992).

III.- DEFINITION OF DOMAINS

The map presented here was constructed from a mosaic of published and unpublished regional data. Whenever the published sources conflicted, the map has been biased to the most recent papers and/or those providing the best supporting evidence. Substage level resolution is generally better for Western Europe than for the rest of the mapped area. With rare exceptions, characterisation of the Early Hauterivian on the Southern Margin of the Tethys is very difficult due to a lack of

reliable biostratigraphic data. This is especially true for the sediments that compose most of the peripheral and inland continental basins referred as "Continental Intercalaire" and/or "Nubian Sandstone" (see age and discussion by KLITZSCH, 1989; LEFRAND & GUIRAUD, 1990). As a consequence, and with respect to the continental formations, the map presented herein is mainly based on a lithological correlation that should better be considered as "Neocomian", even if presumption of Hauterivian age is often the rule.

III.1.- Russian platform (Volga - Ural)

As detailed data on the Early Hauterivian of the Volga - Ural area are scarce, this part of the map has been mainly derived from the compilations by E. AMON and E.J. BARABOSCHKIN. Extension of the marine facies was drawn based on the *Homolomites bojarkensis* ammonite Zone considered as a time equivalent to the Acanthodiscus radiatus ammonite Zone of the Tethyan standard scale (ZAKHAROV & BOGOMOLOV, 1989; SHULGINA, 1989). Nevertheless, a Late Valanginian age cannot be ruled out for the former (KEMPER & JELETZKY, 1979; RAWSON 1981). In any case, a review of the literature evidences a consensus of the majority of the authors that a global regression affected the area around the Valanginian - Hauterivian boundary and is to put in relation to uplifts in Middle Asia. Hence, the fact that shallow marine sediments are limited to the northern part of the Russian platform while continental and non-deposition took place over the Volga region.

III.2.- Moesian platform

I. SANDULESCU and E.J. BARABOSCHKIN having provided separately two different maps, the contours and facies presented herein are the result of a confrontation of those maps with the validation points submitted by the other co-authors.

III.3.- Central and Western Europe

The Danish - Polish furrow correspond to the Polish Lowland where Early Cretaceous deposits are essentially limited to a large sedimentary unit, which is situated between the East European platform and the Variscan platform of the Fore-Sudetic area. The Early Hauterivian stratigraphy and palaeogeography of the Danish - Polish furrow is mainly known from the synthetic maps published by MAREK (1989).

Since Early Valanginian times, the Danish - Polish furrow was connected with the West European seas and the Mediterranean Tethys (KEMPER *et al.*, 1981). The best Early Hauterivian exposure occurs in a brickyard at Waval (Tomaszow Mazowiecki) and the most complete sequences develops in the depocentre (e.g., the Kujavian - Pomeranian swell and adjacent throughs) (MAREK & RACZYNSKA, 1973).

Open marine conditions, similar to that of the Lower Saxony basin, witness to a continuing and expanding transgression that started during Early Valanginian times. The Early Hauterivian sediments are dominated by siltstones and claystones. The greatest accumulations of coarse grain material may be observed in the Kujavian swell and adjacent trough and facies distribution suggesting a northern source of detrital material.

Early Cretaceous palaeogeographic reconstructions of Germany are mainly based on the atlas by SCHOTT *et al.* (1967-69) and ZIEGLER (1988). The Lower Saxony basin expanded considerably during Early Hauterivian times. In the south-eastern part of the basin, extensive transgression took place. The sea was generally shallow, especially during the Amblygonium Zone and, as a consequence, biotopes and facies are variable. On the southern margins of the basin, the rich shallow marine faunas (mainly oysters) occur in limestones containing also iron ore and other coarse clastic material ("Hilskonglomerat") (see KEMPER, 1973 for details).

In the depocentre, full marine conditions prevailed since Early Valanginian times. Deeper water facies are represented by dark clays and shales rich in ammonites. Contrary to Late Valanginian times, oxygenation of the bottom water merely linked to enlarged communication with the North Sea through the Pompekj Swell allowed the development of benthic life such as oysters, crinoids and serpulids.

Early Cretaceous sedimentation in the Western Carpathian (Czech and Slovak Republics) occurred in two basic mega-units, which correspond to the Outer and Central Carpathians. The Outer Carpathians basins were situated in the area of the Palaeo-European Shelf in the foreland of the Bohemian Massif (MICHALÍK, 1993, 1994). The later phases of the Alpine folding in the Late Cretaceous (Central Carpathians) and in the Tertiary (Outer Carpathians) led to complex nappe structure of both units, which became part of the extensive Alpine mountain belt.

Our concern for the Peri-Tethys programme is limited to the Outer Carpathian domain where Early Hauterivian sediments are represented in the Silesian unit (Baska Ridge, Godula basin) and the Pieniny Klippen Belt. The palaeogeographic reconstruction presented here is greatly inspired by the Early Hauterivian map and interpretation of VASICEK *et al.* (1994).

The exclusively marine sediments of the Outer Carpathian basins may be - in a simplified way - divided in two facies. Dark pelites with ironstones nodules are characteristic of the Silesian unit, while light grey marly-lime pelagic sediments develop in the Pieniny Klippen Belt. For both facies cephalopods are the dominant fossil group. Nevertheless, the association of benthic macrofaunas and ammonites in condensed and reworked sediments are a common feature the Baska Ridge, which suggest deposition in an open shelf environment. By contrast, the thick sequence of distal turbidites of the Godula basin assumes hemipelagic sedimentation, while the Pieniny Klippen Belt is characterised by deep-water carbonates that correspond to a pelagic environment.

In north-eastern England, two coastal basins developed full marine conditions throughout the Early Hauterivian (Yorkshire, Lincolnshire). A detailed map and explanatory notes on these areas are to be found in RAWSON (1992), with extensive references. In Southern England, the Weald and Channel basins

correspond to the depocentres of the "Wessex basin complex" *sensu* STONELEY (1982). The Early Hauterivian strata correspond to the lower part of the Weald Clay Formation (ANDERSON, 1973 and 1985; ALLEN & WINBLEDON, 1991). The depositional model of the "Wealden" described at length by ALLEN (1975), as well as this author's view, are now widely recognised. Towards the Eastern borders, the main feature of the Early Hauterivian deposits of the Paris basin is the establishment of open marine environments over an area previously characterised by non-marine sedimentation ("Wealden"). The transgression started in the Late Valanginian times and reached its maximum extension during the deposition of the "Calcaires à Spatangues" (RAT *et al.*, 1987, BULOT, in prep.).

By Early Hauterivian time (Radiatus and Loryi ammonite Zones) most of the south-eastern part of the Paris basin was flooded. The palaeoenvironments correspond to circalittoral conditions characterised by terrigenous sedimentation and soft bottom benthic life (mainly irregular echinoids and endobiont bivalves). To the north-east glaucony might be locally abundant, while to the south-west ooliths dominates. The extent to which the sea penetrated from the French Jura into the Paris basin differs considerably according to authors. Nevertheless, it is generally considered that the "Hauterivian transgression" did not reach Paris, and certainly not entered the departments of Eure, Oise and Seine-Maritime. The limit retained on the map is supported by data of PERNET (1983). North of this limit, the Paris basin maintained fluvio-lacustrine deposition similar to the "Wealden" of the Weald and Channel basins.

Extensive literature was published on the Hauterivian of the Jura as a consequence of early definition of stage in the Neuchâtel area. REMANE *et al.* (1989) published an exhaustive overlook on the subject was and, subsequently, BULOT (1992, 1995) provided updated biostratigraphic interpretation of the Early Hauterivian in the type area. The Early Hauterivian succession of the Jura is subdivided in two very distinctive formations, e.g., the "Marnes bleues d'Hauterive" (Radiatus and Loryi Zones) and the "Pierre Jaune de Neuchâtel" (Nodosoplicatum Zone). As already stated by many authors, the "Marnes bleues" correspond to an open shelf that developed after the drowning of the Jura platform (early Late Valanginian times). The return to carbonate platform conditions did not occurred until the end of the Early Hauterivian (uppermost "Pierre jaune").

Palinspastic reconstruction of the Helvetic shelf is mainly based on the work of TRÜMPY (1980). According to FUNK *et al.* (1993), three distinct palaeogeographic units can be distinguished, e.g., Northern, Middle and Southern Helvetic Realms. With respect to the Early Hauterivian, the stratigraphy of these areas is much better understood since the major thesis work of KUHN (1996). The Early Hauterivian is represented by several lithologies grouped under the name of Kieselkalk Formation (FUNK, 1971; KUHN, 1996). In the Northern and Middle Helvetic Realms (inner and outer shelf), the lowermost Hauterivian (Radiatus Zone) is represented by phosphatic and glauconitic deposits that form part of the "Gemsmättli - Pygurus Complex", a condensed Late Valanginian - Early Hauterivian horizon deposited in open shelf conditions (FUNK *et al.*, 1993; KUHN *et al.*, in press). Towards the Southern Helvetic Realm, this horizon correlates with the "Crioceras beds", a succession of hemipelagic marls and limestones typical of a shelf margin setting.

Detailed Cretaceous palaeogeographic maps of the SE France basin have been published at several occasions, including for the Hauterivian period (ARNAUD-VANNEAU *et al.*, 1982, ARNAUD & LEMOINE, 1993). Recent synthetic studies integrating biostratigraphy, sedimentology and sequence stratigraphy analysis give a new picture of the basin and its margin (ARNAUD *et al.* 1993, BULOT 1995, BULOT *et al.* 1997). The main evolution during Early Hauterivian time consisted in the development of open shelf environments over most of the northern and western margin of the SE France basin. This facies retrogradation was initiated in Late Valanginian times when the Early Valanginian platform drowned. This palaeogeographic change can be recognised over the Gard, Ardèche and Bas Dauphiné areas, and extends north through the whole Jura, connecting the Paris basin with the Tethyan Realm (RAT *et al.*, 1984). By contrast, it should be noted that the Provencal platform and its continuation in Sardinia and Eastern Pyrenees was not affected by the Early Hauterivian changes as carbonate build-ups developed as early as the Radiatus Zone (MASSE, 1976; MASSE & ALLEMANN, 1982).

The map of the Aquitaine basin reproduced here is mainly based on data compiled by Jean-Pierre PLATEL. Fluvio-lacustrine and deltaic "Wealden" sedimentation continued in the west of Aquitaine through during most of the early Early Cretaceous (see also KIEKEN, 1974). The Hauterivian is represented by continental silty claystones and shallow marine sandstone in the Parentis basin. To the west, continental dark clays with Charophytes are known in the Pau basin.

Deltaic and fluvio-lacustrine "Wealden" sedimentation occupies large areas in Northern (Cantabric) and Eastern (Iberic) Spain. Marine sediments are restricted to the Maestrazgo, Southern Spain (Betic) and the onshore Portuguese basins. The Basque - Cantabrian basin has been subject to extensive work due to the initiation and development of the Basque segment of the passive North Iberian resulting from the early Early Cretaceous rifting phase (RAT, 1988; UCHUPI, 1988 with extensive references). The Early Hauterivian is represented by the lower part of the Vega de Pas Formation. Sediments are dominated by fluvio-lacustrine sandstone and mudstone (see details *in* PUJALTE, 1981 and 1982). Age and interregional correlations are mainly based on freshwater ostracods and charophytes (BRENNER, 1976). Comparatively, the Iberian basins have been the object of limited investigations. Age determination between the Valanginian and the Early Hauterivian is still to be ascertained. Palaeoenvironments range from fluvio-deltaic to brackish but timing is unknown (for details see MAS *et al.*, 1982; PÉREZ DEL CAMPO & ZAVALA, 1982; VILAS, ALONSO *et al.*, 1982 and VILAS, MAS *et al.*,1982).

Due to marine deposition, the Early Hauterivian palaeogeography of the Maestrazgo, Betic and Portuguese onshore basins is much better understood, but too complex to be detailed herein. The palaeoenvironments

represented on the map are directly inherited from the synthetic papers of CANÉROT *et al.* (1982), GARCÍA-HERNÁNDEZ *et al.* (1982) and REY (1979). The reader is referred to these papers for further details.

III.4.- Maghreb

The Early Hauterivian Tethyan shoreline can be traced from Morocco to Libya with some confidence. Landward, a vast north flooding fluvio-lacustrine system existed covering much of what is now the Saharian Region ("Continental Intercalaire" and equivalent strata of LEFRANC & GUIRAUD, 1990). GUIRAUD (1973) showed that the base of the Cretaceous is marked by the wide development of a continental realm ("Land of Idrissides"), which cover a large part of Morocco, the Algerian High Plateaux and the Saharian Atlas. South of this peninsula, terrestrial, mostly fluvial, deposits developed during most of the Early Cretaceous (mainly Hauterivian, e.g., MONBARON, 1979).

Marine sedimentation was therefore limited to the Atlantic basin (West Morocco) and part of the Rift foreland basin (North Morocco), where carbonate reefs complex developed during the Late Valanginian and Early Hauterivian (CANÉROT *et al.*, 1986; REY *et al.*, 1988). Similar conditions prevailed also in the northern part of the Pretellian Zone of northern Algeria. It should be noted that on the southern border of these areas, "Wealden" deltaic to coastal marine deposition may intergrade in the carbonate complex (CANÉROT *et al.*, 1986 with references).

In Central and Eastern Algeria Sahara, the Oued Mya basin is the type-area of the "Continental Intercalaire". "Neocomian/Barremian" continental detrital formations are represented by fine-grained sandstones and clays at the base; and coarser-grained sandstones that grade into the gravely and conglomeratic "grès à dragées" at the top. Early Hauterivian is assumed to be represented, though its range is unknown.

As suggested by MATEER *et al.* (1992), a broad correlation of non-marine Cretaceous strata is possible from western Algerian to Libya. According to LEFRANC & GUIRAUD (1990), deposits similar to the "Continental intercalaire", occurs in the west and southwest of the Algerian Sahara (Gouara, Touat and Tidikelt regions). In northeast Hoggart, some part of the "Serouenout Series" may also correlate with the "Continental Intercalaire". Continental deposits of Hauterivian age *sensu lato* have also been reported from the South Tunisian Sahara (CANÉROT *et al.*, 1986), Tripolitania and most of the main Libyan basins (LEFRAND & GUIRAUD, 1990).

In central Tunisia, the marine Hauterivian is represented by neritic to hemipelagic shales, siltstones, turbidites and limestones (for details see BEN FERJANI *et al.*, 1990; M'RABET *et al.*, 1995). Well-dated Early Hauterivian is known from the Jebel Oust and these locality stands as a reference area for the Tunisian Trough (MEMMI, 1989). Elsewhere, lateral facies changes are rapid and render correlations difficult (SOUQUET *et al.*, 1997). Early Hauterivian hemipelagic sediments (glauconitic limestone and dark shales) have been reported in the north-eastern part of Cyrenaica (Libya), while over the rest of Northern Libya deltaic to coastal marine conditions prevailed during Berriasian to Hauterivian times (REYNOLDS *et al.*, 1997 with references).

III.5.- Ethiopia - Egypt

The above mentioned facies belt can be traced eastwards to the Nile delta where sands and shales deposited during the Neocomian (including the Hauterivian) over most of Northern and North-West Egypt (SCHRANK, 1992; REYNOLDS *et al.*, 1997). Nevertheless, around Matruh and Northern Sinai, deeper basins with predominance of shale deposition witness to full marine conditions.

To the south, increasing continental influence in the coastal sands assume the transition to the continental and fluvio-deltaic "Nubian Sandstone" of the Egyptian basin that exhibits facies evolution similar to the Algerian Sahara (see discussion *in* LEFRANC & GUIRAUD, 1990). Age and stratigraphic relations between the different continental facies registered as "Nubian Sandstone" has been elucidated (KLITZSCH, 1990; WYCISK, 1990, 1991; WYCISK *et al.*, 1990). An extensive overview on the Early Cretaceous non-marine deposits of Southern Egypt and Sudan with full references is to be found in the synthetic paper by MATEER *et al.* (1992).

As a rule, solid evidence of Early Hauterivian sediments in the Horn of Africa (Ethiopia, Somalia) are scattered. Nevertheless, BOSELLINI (1989) and LUGER *et al.* (1994) provide the more consistent synthetic papers on the sedimentary evolution of this region based on published and unpublished data. According to these authors, in Northern Somalia, the erosion of early Lower Cretaceous sediments including Hauterivian, is to be related relation to widespread uplift and severe block tectonic movements. BOSELLINI (1989) suggested that this process took place in two phases, as indicated by the presence of unconformity-bounded Wealden-type sediments between the Jurassic and the Aptian marine deposits. By the same time, Central Somalia underwent continued subsidence with evaporitic sedimentation (e.g., the "Main gypsum") in Ogaden and marine deposits in the Mudugh basin. To the south, continental and evaporitic deposits filled in the Mandera Lugh basin.

III.6.- Syria - Levant

The extension of the "Nubian sandstone" into the Middle East has been established long since. During the Cretaceous, north-eastern Africa and the Middle East formed a continuum in the absence of the Red Sea. The southern margin of the Mediterranean Tethys extended from the north coast of Africa into Israel, Lebanon, Syria, Iraq, Northern Saudi Arabia, Oman and coastal Somalia.

In Israel, Hauterivian marine shales dated by ammonites and ostracods have been recognised in wells on the Coastal Plain (ROSENFELD & RAAB, 1980). Offshore and onshore, this highly detritic sequence intergrades in the deltaic siltstone of the Helez Formation (HIRSCH, 1990). East and southwards, continental sandstones of the Hatira Formation developed over Galilee and Neguev.

Regional uplift leading to regression occurred widely in the Middle East between the Late Jurassic and Early Cretaceous. This results in either non-deposition of any Early Cretaceous sequence or severe erosion of older strata. It seems that the Palmyrides region of Syria, South-western Syria and parts of Western Iraq remained emerged during most of the Early Cretaceous. Nevertheless, deltaic sandstones and conglomerates deposited in North-eastern Syria and fluvial sandstones and shales in Jordan are thought to be "Neocomian" (REYNOLDS *et al.*, 1997 with references).

III.7.- Arabian plate

In North and Central Saudi Arabia, shallow marine shelf deposits covered most of the early Valanginian platforms. Similar conditions extended towards the United Arabs Emirate and Interior Oman. The shelf edge of the Arabian platform was therefore under deep marine conditions (shales and limestones).

In Yemen, Rift failure at the end of the Jurassic was ended by subsidence that resulted in a basin fill phase and continental sandy conglomerates deposited all over the area during the Cretaceous. There is strong disagreement about the age of these deposits and very little palaeontological support to the age proposed. As Hauterivian age cannot be ruled out, the co-ordinator has made the choice to represent continental facies over western Yemen (see also discussions *in* MATEER *et al.*, 1992).

IV.- DESCRIPTION OF DOMAINS

IV.1.- Palaeoenvironments

Compared with their extension during the early Valanginian and Barremian to early Aptian, the extension of early Hauterivian carbonate platforms is very restricted. As already stressed out by MASSE *et al.* (1995), Peri-Mediterranean areas such as SE France, Swiss Jura, Portugal, Spain as well as some parts of Eastern Europe, North Africa and the Western Pontides, the abrupt collapse of carbonate settings took place as early as the Late Valanginian and carried out during most of the early Hauterivian.

As a consequence of the palaeogeographic (and most probably palaeoclimatic) changes exposed above, during early Hauterivian times, drowned platform environments are widespread over large parts of the north-Tethyan margin. They usually correspond to circalittoral environments in which a large spectrum of endobiont bivalves and irregular echinoids developed. On the platform margins, phosphoritic and glauconitic deposition is almost the rule and the sequences are affected by non deposition, reworking and condensation.

IV.2.- Facies

The Hauterivian fluvio-lacustrine facies is often refereed to as the Wealdian. Everywhere, this facies is represented by sandstones and clastic shales. Large to medium scale planar crossbedding is the dominant sedimentary structure and is taken to indicate braided or low sinuosity streams in alluvial plains. Several authors (see ZIEGLER *et al.*, 1987 for examples) have also assumed flooded activity has also been assumed. The facies is extensively distributed in Western Europe and characterise large parts of the Iberian marginal basins, the Parentis and Aquitanian basins, and the Anglo-Paris basin (see fig. 6 *in* UCHUPI, 1988). Equivalent facies in North Africa and the Middle East are known as the "continental intercalaire" and the "Nubian sandstone" (see MATEER *et al.*, 1992 for maps and details).

The extension of evaporitic facies is very limited during early Hauterivian times. An exhaustive review of Early Cretaceous evaporitic deposits was published by HALLAM (1984). The main areas of deposition are North-eastern Somalia ("Main Gypsum", BOSELLINI, 1989) and the former USSR, from Western Caucasus to Kopet Dagh (VAKHRAMEEV, 1987). In both areas, evaporites are associated with continental red beds.

Two main facies type characterised the restricted early Hauterivian carbonate platform. Calcareous shallow-marine, biogenic mounds are quite widely distributed and developed mainly in SE Spain, Crimea, Moesia and along the present coast of western North Africa (CANÉROT *et al.*, 1986; MASSE *et al.*, 1995). Rudist build-ups are poorly represented and restricted to the upper part of the substage. They have been reported from Spain, Provence (SE France), Jura and Sardinia (MASSE & ALLEMAN, 1982; MASSE *et al.*, 1995).

Deeper water carbonates are mainly represented by megarythms of marl-limestone alternations. These micritic mudstones have a high carbonate content up that represent up to 80% of the bulk rock. The sediment is a consolidated nannofossil ooze usually very bioturbated with low TOC content. They are best expressed in SE France but equivalent facies are also known in many other areas of the North Tethyan margin.

V.- CONCLUSIONS

Three main trends in palaeogeographic changes can be recognise over the studied area. In Western Europe, a generalised transgression initialised during early to Late Valanginian times characterises the epicontinental basins of NW Europe. In the meanwhile, most platforms and shelf areas are drowned and extensive hemipelagic deposition took place on the Northern margin of the Tethys. However, this palaeogeographic changes reflect an opening of the marine environments rather than a deepening of the shelf areas.

On the contrary, in Eastern Europe, early Hauterivian times seem to be more marked by a regressive trend. Most probably, deposition of continental strata (subsequently eroded) covered large areas of the Urals and East European platforms. Marine conditions prevailed in the north-eastern border of the Tethys, where limited carbonate build-ups developed in the Dobrogea and Crimea.

Southern Tethys sedimentary basins are characterised by terrigenous and detritic sedimentation linked to uplifts of the African Craton. In general, early Hauterivian

is not distinctive from the rest of the Neocomian - Barremian sedimentary cycle. Limited marine carbonate build-ups developed during the early Hauterivian on the western part of the area considered (Morocco, Algeria), in contrast to the Arabian platform, already drowned by Late Valanginian times.

Geodynamic control, climatic changes and eustatic sea level rise seem to have played major roles in the settlement of the Early Hauterivian palaeogeography and implied a well marked variation of their respective impact on the various regions. They will probably open new paths to research in a near future.

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Acknowledgements

The map and explanatory notes presented here are the result of the joint efforts of the regional and map co-ordinators listed above. Although as much updated data as possible were synthesised, the map co-ordinator takes on the entire responsibility for the final choice of contours and palaeoenvironments. As a general rule, precise biostratigraphy prevailed all over this work and lithostratigraphy was chosen as an alternative (see reference list) only when data was missing.

This work would not have been achieved without the fruitful discussions with colleagues during the meetings, congresses and excursions of IGCP Project 262 (1989-1993) and 362 (1993-1998). Quite often, workers out of the scope of the Peri-Tethys Programme facilitated the access to published data. Among them special thanks are due to Prof. D.J. Batten (Wales University, UK) for his tremendous help in the collection of references on the non-marine Hauterivian, and to Dr. Z. Lewy (Geological Survey, Jerusalem) for providing numerous papers on the Middle East. Grateful acknowledgements are expressed to Société de Secours des Amis des Sciences for the grant (1996-1997) provided to support the co-ordinator's work.

13.- EARLY APTIAN (112 - 114 Ma)

J.-P. MASSE[1]

I.- MAIN FEATURES

The duration and boundaries of the Aptian, based on radiochronologic data, have evolved since they were first proposed :

CASEY (1964)	112-106 Ma (6 Ma)
VAN HINTE (1976)	115-108 Ma (7 Ma)
HARLAND *et al.* (1982), PALMER (1983)	119-113 Ma (6 Ma)
KENT & GRADSTEIN (1985), HAQ *et al.* (1987)	113-108 Ma (5 Ma)
ODIN & ODIN (1990)	114-108 Ma (6 Ma)
HARLAND *et al.* (1990)	124-112 Ma (9 Ma)

Furthermore, recent studies performed in Israel tend to confirm the HARLAND *et al.* (1990) dates, and proposed an Early/Late Aptian boundary older than 118 Ma, giving a duration for the Early Aptian of about 4-5 Ma (GVIRTZMAN *et al.*, 1996).

On the magnetostratigraphic scale, the Early Aptian is asked to coincide with the base of magnetic chron MO(R) (ERBA, 1996), whereas palaeontologic data to calibrate this event, are poor. The "ISEA subchron" detected in the Apennines and in Israel, in the C34N normal polarity interval, is Gargasian in age and dated at 118 Ma (GVIRTZMAN *et al.*, 1996).

In its historical-type locality at the Bèdoule (SE France) the "Bedoulian" (TOUCAS, 1888 *in* BUSNARDO, 1984) was divided by the latter into seven ammonite zones, though a recent revision of the stratotype shows the existence of four zones, (MOULLADE *et al.*, 1998) shown on table 13.1, including correlation with ammonite zonations proposed by various workers for Peri-Tethyan areas.

S.E. France (1)	La Bèdoule (2)	S. England (3)	Turkmenistan (4)	
Bowerbanki	Furcata	Bowerbanki	Furcata	
Grandis	Grandis	Deshayesi	Deshayesi	Late Bedoulian
Hambrovi	Hambrovi			
Matheroni	Deshayesi			
	Weissi	Forbesi	Weissi	Early Bedoulian
Consobrinus	Tuarkyricus	Fissicostatus	Tuarkyricus	

Table 13.1 : Ammonite biozonations for the Early Aptian in Peri-Tethyan domain.
1- BUSNARDO (1984), 2- MOULLADE *et al.* (1998), 3- CASEY (1964), 4- BOGDANOVA (1971) *in* DELANOY (1995).

Planktonic foraminifera are not reliable markers for the substage boundaries: *Praehedhergella kuznetsovae* appears in the latest Barremian (Sarasini Zone) while *Schackoina cabri* appears in the Deshayesi Zone; nevertheless, *Blowiella blowi* starts precisely at the Early/Late Bedoulian boundary. For calcareous nannofossils, *Rucinolithus irregularis* points out to the Barremian - Aptian boundary (BIRKELUND *et al.*, 1990), while *Eprolithus floralis* appears at the base of the Furcata Zone (MOULLADE *et al.*, 1998). As a consequence, these results only partly agree with those of ERBA (1996).

The range charts of spores, pollens and dinoflagellates show that some taxa seem to be useful for inter-regional correlations (JARDINÉ *et al.*, 1984).

Many species of faunas and floras belonging to carbonate platform communities are well known as chronologically significant Early Aptian markers. This is the case for Orbitolinids: especially the *Palorbitolina lenticularis* (Blumenbach) - *Praeorbitolina cormyi* Schroeder couple. Many Dictyoconids have a restricted regional extension and therefore a limited value for inter-

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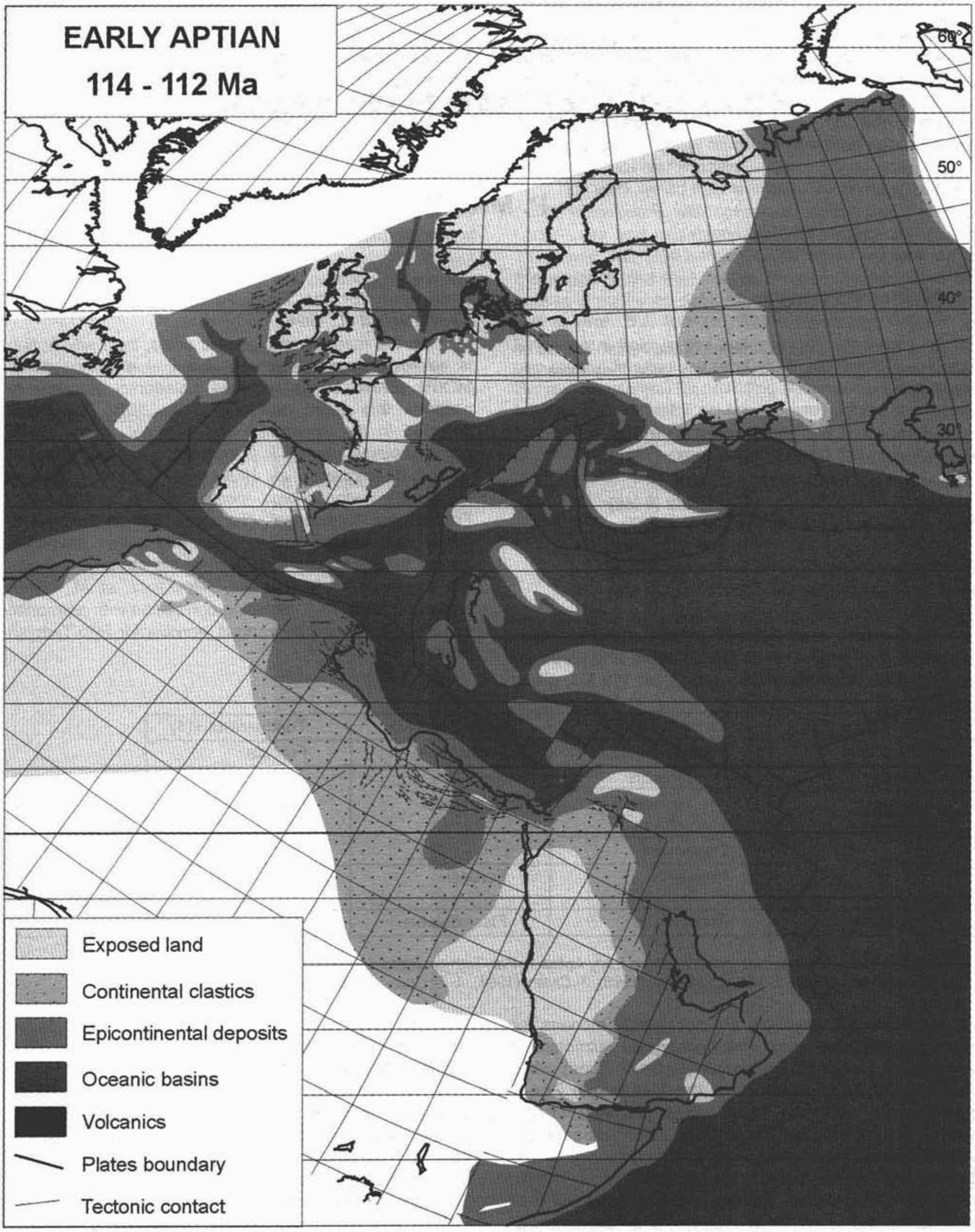

Fig. 13.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Aptian.

regional correlations, except: *Palaeodictyoconus arabicus* (Henson) and *Rectodictyoconus giganteus* Schroeder known from Western Asia to Western Europe (PELISSIÉ *et al.*, 1982). Among Rudists, the Early Aptian is distinguished through the development of Caprinids: *Offneria, Caprina, Praecaprina* from the Mediterranean regions and the Middle East.

Among calcareous algae, many species of dasycladales are limited to the Early Aptian, which appears to be a golden age for this group (MASSE, 1993).

At a global scale, the Aptian - Albian major marine inundation on the continental platforms was more extensive than the "Neocomian transgression" (MATSUMOTO, 1980). This tendency is well expressed in Europe (TYSON & FUNNELL, 1990), Iberia (CANÉROT *et al.*, 1982; ARIAS *et al.*, 1987; FERNANDEZ MENDIOLA, 1989), North Africa (MASSE, 1984), and South Arabia (ROGER *et al.*, 1989). Nevertheless, regressive events are known punctually, as in Morocco (BEHRENS & SIEHL, 1982).

The dominant transgressive trend is commonly interpreted as the result of a global sea level rise. The global sea-level curves from WATTS & STECKLER (1979) and WATTS (1982) *in* WILLIAMS (1988), suggest a nearly 10 m rise during this period. Based on the Russian platform section used as a "stable frame of reference", SAHAGIAN & HOLLAND (1991) calculated a sea level rise of about 25 m.

The global sea level curve of HAQ *et al.* (1987) shows a sharp sea level fall during the earliest Aptian (Consobrinus - Matheroni Zones) of about 60 m, then followed by a rise of the same amplitude. As mentioned by CHRISTIE-BLICK (1990) "until a good deal of further work is completed, amplitudes of sea level changes, for the Cretaceous in general, indicated by the HAQ *et al.* (1987) curve, must be regarded as conjectural and possibly as much as two or three times too large".

Except for the very fast rate of sea level changes (nearly 100 m/Ma) suggested by HAQ *et al.* (1987), rates provided from other models (i.e. 5 to 10 m/Ma) are consistent with the rates based on calculations of the changes in volume and length of the global mid-ocean ridge system (PITMAN & GOLOVCHENKO, 1983). However, non-eustatic hypothesis for Aptian transgressions have been poorly explored. The replacement of terrigenous clastic by carbonates, recorded in many regions, without significant bathymetric changes, could suggest an important climatic control. Whatever the possibility of a sea level rise, the corresponding assumed rates are comparable to those of subsidence, although sedimentation ones could have been significantly higher.

According to BARRON *et al.* (1980), KEMPER (1987) and FRAKES & FRANCIS (1990), Early Aptian was a warm period. Postulated greenhouse conditions imply elevated CO_2 content in the atmosphere (WEISSERT, 1989). Moreover, WALKER (1986) also proposes a reduction of the atmospheric oxygen.

Between 130 and 110 Ma, the earth radiation budget shows an exceptionally high global value, which is correlated with a low percentage of land area to the whole earth surface (BARRON *et al.*, 1980). Such conditions imply:

- a low (30% of the modern value) equator pole oceanic surface temperature gradient (LLOYD, 1982), therefore the zonal mean wind speed in the westerly belt would have been reduced,
- tropical circulation with a strong monsoonal characteristic along the margins of the Mesogée and in the Central Atlantic.

Aptian continental fossil floras from Eastern Asia are indicative of relatively high temperature (by comparison with Albian and Barremian ones), well illustrated by an elevated Cycadophyte quotient (KRASSILOV, 1973). VAKHRAMEEV (1964) hypothesised the same evolution for Europe and North America. Even if the Early Aptian global climate seems to fit in with the "Cretaceous warm climate" model, the existence of freezing conditions, allowing the formation of polar ice caps in the polar zones during this period, has been suggested (FRAKES & FRANCIS, 1990).

Assuming that seasonal surface air temperature ranges over land was probably stronger than today, i.e. continentality would have been more severe, a pattern correlated to weak circulation (LLOYD, 1982), geological data fit in with the palaeoclimate reconstructions. .

Oceanographic changes on a time scale of 10-100 Ka documented from the deep-sea sedimentary record and interpreted as the expression of astronomical controls (eccentricity and precessional forcing Milankovitch rhythms) (HERBERT & FISHER, 1986), suggest coeval climatic modifications on the continents.

Atmospheric circulation is assumed to drive the surface oceanic circulation. Consequently, a westward flowing surface current circling the globe in the intertropical zone was proposed by LUYENDYK *et al.* (1972), GORDON (1973), BERGGREN & HOLLISTER (1974) and LLOYD (1982) whereas BARRON & PETERSON (1989) postulated an eastward flow.

The latitudinal range of carbonate platforms, with the same palaeoclimatic meaning as modern coral reefs, is between 35° N and S (PHILIP, MASSE & CAMOIN, 1996). This kind of distribution is a good illustration of the high latitude drift of warm seas by comparison to the present day.

Oceanic vertical temperature gradients were probably considerably low compared to the Quaternary gradients. Sea surface palaeotemperatures calculated after oxygen isotopes ($\delta^{18}O$) at La Bédoule are 19°-22°C (KHUNT *et al.*, 1998) while for the Mid-Cretaceous tropics they fluctuate between 25°C (SELLWOOD *et al.*, 1994) up to 30-31°C (NORRIS *et al.*, 1998; FLUTEAU 1999).

A peculiar feature of the hydrologic structure of the ocean is in that owing to elevated mean ocean salinity (>43‰) thermohaline circulation was different from that of today. The regions with high water densities, potential sites of intermediate or deep water formation, were located near the poles and the tropics, that is to say from the Atlantic side of Morocco to the eastern Pontides. For HAY *et al.* (1997), high salinities promoted active thermohaline circulation and particularly more intensive intermediate water formation. Intermediate water is regarded having high concentrations of nutrients and is the major source for nutrient rich upwelled water, leading both to surface high productivity and subsurface oxygen depletion required to the formation of black shales.

This model contradicts the "sluggish circulation" model postulated after low pole-equator thermal gradients (JENKYNS, 1980). Such a model provides a clue for understanding black shale formation in the Peri-Tethyan domains: Goguel and Selli organic rich beds (ARTHUR *et al.*, 1990) pointing out to the role of OM preservation. Nevertheless, increasing production (high nutrient peaks) coupled with continental derived products is also considered prominent factors for the production of organic rich sediments (ARTHUR *et al.*, 1990). The high potential of OM preservation linked to the ocean hydrologic structure was enhanced by the

drowing down of oxygen, which has been linked to large fluxes of sulphide production by rapid hydrothermal activity (WALKER, 1986).

Geochemical data, highly significant concerning the oceanic global environment, have been analysed in pelagic deposits by RENARD (1985), as concerns the Early Aptian. This period is characterised by the following parameters :

- Mg^{++} and Sr^{++} drop sharply allowing to consider the Early Aptian as the end point of the "Early Cretaceous geochemical cycle". These modifications are responsible for a change from an aragonite regime known during the Barremian - Early Aptian to a calcite regime in the Late Aptian.

- $\delta^{13}C$ is high (i.e. in pelagic carbonates), its increasing rate is remarkably high during the Early Aptian. This phenomenon coincides with the so-called OAE 1a (oceanic anoxic event) (JENKYNS, 1980) or E1 event from the Atlantic (GRACIANSKY *et al.*, 1982). Elevation of $\delta^{13}C$ is connected with the organic carbon burial (depleted in C^{13}) in relatively deep-water environments (SCHOLLE & ARTHUR, 1977 *in* ARTHUR *et al.*, 1990). Furthermore, organic carbon burial rate is thought to have been also very high.

-$\delta^{34}S$ of marine evaporites is depleted (CLAYPOOL *et al.*, 1980 *in* ARTHUR *et al.*, 1990).

A peak in $\delta^{13}C$ is recorded in various areas of the Tethyan domain, both in pelagic and shallow water carbonates, this positive excursion is coeval with the Hambrovi and Grandis subzones of the Deshayesi Zone (KHUNT *et al.*, 1998; MOULLADE *et al.*, 1998; MASSE *et al.*, 1999).

On carbonate platforms, European and Africa - Arabic faunas are well defined by the occurrence of characteristic rudist genera (MASSE, 1985): *Glossomyophorus* is known from Somalia to Italy, and species, especially from the *Offneria* genus.

Many species of foraminifera belonging to the genera: *Lenticulina, Dentalina, Gaudryina, Ammobaculites, Verneuilinina* are known from the Boreal and Mesogean realm (BARTENSTEIN, 1979; MAGNIEZ-JANIN *et al.*, 1984). Extra Mesogean foraminifera are mainly represented by relatively "deep water" Mesogean forms, their biogeographic meaning is therefore unclear. This wide biogeographic distribution allows using many of these species as potential "world-wide biostratigraphic markers" (see BARTENSTEIN, 1979; ASCOLI, 1976; MAGNIEZ-JANIN *et al.*, 1984). The planktonic foraminifers also show a wide biogeographical distribution: Mesogean and Boreal forms are near identical (BOLLI, 1959; LONGORIA, 1974; ASCOLI, 1976; MAGNIEZ-JANIN *et al.*, 1984; ERBA & QUADRIO, 1987).

The "Tethyan realm" is clearly expressed by ostracod distribution with a latitudinal division corresponding with the European and African provinces. The Middle East and North Africa are characterised by the *Antepaijenborchella* fauna (BABINOT & COLIN, 1988).

Among belemnites, the *Belemnopsidae* are known from the Mesogea and the boreal domain, while the *Duvaliidae* (*Duvalia*) are typical inhabitants of the Mesogea (COMBEMOREL, 1984). As noticed by STEVENS (1963), the provincialism is less pronounced during the Aptian than during the Neocomian. Early Aptian is marked by the spreading of many Tethyan genera in the temperate adjacent regions. Similarly, most of Mesogean ammonites are known in the boreal domain, whereas some endemism is reported from Russia.

The distribution of continental floras allows to separate the North Gondwana and the South Laurasia floristic domains (PONS & VOZENIN-SERRA, 1984).

II.- STRUCTURAL SETTING AND KINEMATICS

Palaeogeographical reconstructions of Peri-Tethyan areas and, especially, the question of the Cretaceous events between the African - Adriatic and the Eurasian plates are still contradictory. Moreover, the number of plates acknowledged by various authors varies from 2 to 6 (CHANNEL *et al.*, 1979; FRISCH, 1981; VAN DER LINDEN, 1985; SAVOSTIN *et al.*, 1986; RICOU, 1996)

The reconstruction is based on assumption that the European part of the Laurasia includes blocks and microplates: Iberia and Preapulian domain (MASSE *et al.*, 1993). The Apulian domain (i.e. Adria) is separated from Africa by the Eastern Mediterranean basin (RICOU, 1996). The eastern Tethys embayment (flanking the eastern part of the Mediterranean seuil) pertains to the western part of the so-called oceanic K transit plate (RICOU, 1996).

The ridges which separate the K-Transit plate from Africa are reconstructed from the peri-Arabic ophiolites, now obducted; the corresponding directions of the spreading axes for the Taurus and Oman areas require the existence of a third branch. This branch separates an Apulian plate from a Transit plate and is somehow inherited from a Late Jurassic major transfer fault. Its location is chosen to account for the Cretaceous break-up and ophiolite formation in Iran (RICOU, 1996).

The position of the palaeomagnetic pole is taken from BESSE & COURTILLOT (1991). The diverging boundary between Africa and the Apulia microplate accounts for the opening of the eastern Mediterranean ocean. Whether the initial opening of this oceanic domain is pre-Cretaceous (ROBERTSON & DIXON, 1984; ROBERTSON *et al.*, 1996), or Early Cretaceous in age (RICOU, 1996) is still debated. Similarly, the age of the Pamphylian oceanic segment, which separates the Arabian promontory from the Taurus - Menderes block, is poorly documented, ROBERTSON & DIXON (1984) and ROBERTSON (1998) regard this zone as the eastern part of their Levantine oceanic basin opened during the Triassic, while POISSON (this volume) postulates a Cretaceous age. The Serbo-Macedonian - Rhodope and the Pontides - Caucasus margins are considered either mainly passive (SENGOR *et al.*, 1984) or active (ROBERTSON & DIXON, 1984; ROBERTSON 1998; MASSE *et al.*, 1993). Moreover, the Black Sea opening and the existence of a Pontides - Caucasus volcanic arc system are also controversial. ROBERTSON & DIXON (1984) assumed that the opening of the Black Sea was Late Cretaceous whereas MASSE *et al.* (1993) postulated on Early Cretaceous age.

Field data from the Black Sea coast of Turkey shows that volcanic arc systems started in Cenomanian times (CHARLES & FLANDRIN, 1929).

III.- DEFINITION OF DOMAINS

III.1.- The Russian platform

Early Aptian records a significant basin enlargement in Moscow and the Caucasus regions due to marine transgressions. The basin is still connected with the Arctic domain but connections with NW Europe through the Polish straight are interrupted. The Baltic and Ukrainian shields are emerged; the Ural uplift represents a barrier between the Russian and West Siberian basins. Fine terrigenous sediments dominate over carbonates, with a clear fining trend from the emerged land to the central part of the basin. Early Aptian times record a wide opening of the Moscow region to the Mediterranean domain, allowing a migration of the Tethyan ammonite fauna to the boreal and arctic regions (BARABOSHKIN, 1997). The best sections of the Russian platform are found in the Middle Volga River valley between Ulyanovsk and Saratov in the Simbirsk Syneclise. The Forbesi, Deshayesi, Grandis and Bowerbanki Zones were identified with some *Australiceras*. Sediments consist of glauconite - quartz sandstones and grey clays in the lower part, black shales and silts in the upper part of the sections (BARABOSHKIN, 1998).

III.2.- Moesia, Crimea, Caucasus, Turkmenistan

Coastal clastics are found in Southern Dobrogea (Romania) lining the Capidava - Ovidiu fault (SANDULESCU, 1984; DRAGASTAN *et al.*, 1998); southwards, these sediments pass to platform carbonates, which develop in NE Bulgaria (Ruse platform). The Moesian platform and the Central basin is also fault bounded (IVANOV *et al.*, 1997). In Bulgaria, conditions tend to continue the patterns of the Barremian.

In addition to their development in the region of Veliko Tarnovo - Lovec, Urgonian facies developed over the eastern and western Fore-Balkan, western Srednogorie, western Balkanides, and also in the regions of Vidin and Orjahovo. This zone was bordered by a wide band of shallow-water clastics and argillaceous sediments. Marine littoral sediments are found to the north of Kotel. Only a relict remained of the old sedimentary basin, with a fill of sandy-marly deposits (TCHOUMATCHENKO *et al.*, 1990).

The SW part of the Rhodope massif (Northern Greece and Bulgaria) is considered as a block that migrated away form Gondwana and was accreted to the active margin of Eurasia during the Mid Cretaceous: this block named Drama is somewhat comparable to the Turkish Kirsehir one (RICOU, 1996).

In Crimea, Early Aptian sediments are only recorded in the central and eastern parts of the peninsula where they are represented by marls with sandy beds containing ammonites (*Deshayesites*) (NIKOLOV, 1987).

In the Great Caucasus, ammonite-bearing marls predominate; similar facies are found in Georgia (Kutaissi, Abhkhazia). There is a well-marked north-south trend in thickness and facies changes. Thin glauconitic sands are found in the north while thick marls are found in Central Caucasus (NIKOLOV, 1987). In this region, tectonic instability is illustrated by the presence of Barremian klippes made of Urgonian limestones; the Dibrarian klippes from Altyagach, Azerbaidjan (YANIN, 1990), are reworked in Aptian marls containing belemnites. Notwithstanding, repeated assumptions concerning an Aptian age of the volcano-sedimentary systems of the Lesser Caucasus (see NIKOLOV, 1987) and biostratigraphic data provided from Nakichevan (southern Armenia) show that the volcanic-arc sediments post-date *Mortoniceras* bearing beds (Late Albian) (BONNET & BONNET, 1947).

Marly facies extend to the Turanian platform (Turkmenistan) and pass laterally to continental red-beds in northern Tuarkyr and the Amou - Daryan syneclise (Boukhara) where fluvio-deltaic sediments are found (PROSOROVSKY, 1990). Aptian times are characterised by the drowning of pre-existing continental uplift (south Caspian Sea and Mangyshlak) a strong facies homogenisation (ammonite bearing marls), whereas in the Bolshoi Balkan large benthic foraminifera (*Palorbitolina* and *Balkhania*) indicate shallow water conditions (SCHROEDER & DE LAPPARENT, 1967).

III.3.- Western Europe

The North Sea basin was widely open towards the north (Rockall - Faeroes trough), the Central Graben and its northward extend, the Viking Graben, were in the course of passive infilling. The Grampian - Pennine and London - Brabant massif were united, interrupting the connection between the North Sea domain and the Channel-Paris marine seaway. The Sole Pit Inversion was uplifted. Aptian time is looked as the overturning phase of the tectono-sedimentary regime resulting in a distinctive segregation between the North Sea, Atlantic Rift and Barents Sea domains. North Sea began a transition towards a more stable regime of epicontinental subsidence. The reddish calcareous claystones of the Rodby Formation were deposited throughout the North Sea (DORÉ, 1991; BRUN & TRON, 1993). Condensed deposits are found over crests of fault-blocks (RUFFELL, 1991).

Marls and sandstones (Atherfied Clay and overlying Hythe beds) developed in Southern England in the Weald basin the geometry of which is controlled by faults (in particular the Hogs Back - London platform fault and Folkstone fault on the northern and eastern sides of the basin). The Hythe sandstones tend to be transgressive over the underlying clays, the sediment source probably lay to the west (RUFFELL, 1992). In the offshore of SW England, Early Aptian beds are frequently missing: Albian sediments overlay pre-Cretaceous rocks. Notwithstanding, a marine regime is hypothesised for the western English Channel between the Cornubian and the Armorican highs; carbonaceous clays are documented from the South Western Approaches basin (ODP site 402A) (LOTT *et al.*, 1980). Speeton (Yorkshire) provides an outcrop view of the typical North Sea succession: Aptian marls overlay Barremian marls and limestones, marls are also found in the Danish Central Trough of Southern North Sea (RUFFELL, 1991).

The Holland Marl represents the southern extension to Netherlands of this marly regime, which tends to rest on a Barremian - Aptian erosion surface. In the lower Saxony basin of Germany, the Bodei clays are overlaid by black shales (Fish shales). In the Polish Trough glauconitic sands became widespread but, as in the Barremian, connections with Tethys were interrupted (TYSON & FUNNEL, 1990).

In the Paris basin, marly facies are predominant and are mainly represented by the Plicatula marls (Late Bedoulian). This episode is thought to record the connection between the Alpine and the English Channel domains (CORROY, 1925). Significant erosions of antecedent platform limestones and the age of the Aptian transgression in the French Jura, (Furcata Zone) (CLAVEL *et al.*, 1995) makes difficult the existence of a wide connection between the Paris basin and the Alpine domain during the Early Aptian.

In contrast to the foregoing regions, Southern France and Northern Spain are characterised by the development of Urgonian carbonate platforms, which run to Switzerland and Sardinia (MASSE *et al.*, 1993). A typical feature of SE France is the horseshoe pattern of Urgonian platforms surrounding the Vocontian pelagic basin (and its associated hemipelagic belt) (ARNAUD-VANNEAU *et al.*, 1979, MASSE & PHILIP, 1980). The Aquitaine - Pyrenées intrashelf basin is connected to the South Provence hemipelagic basin (where the historical La Bedoule stratotype is situated). This depression corresponding with the Sainte-Suzanne marls connects with the Catalonia marginal basin, though uplifting is hypothesised to the west, which precludes a communication with the Bay of Biscay.

Iberia is regarded detached from Europe along the Bay of Biscay longitudinal faults. On the northern side of the Bay, mixed carbonate terrigenous sediments are found along the Armorican margin, including isolated carbonate platforms off Brittany (Meriadzek). On the southern side, the Basco-Cantabria Urgonian platforms flank the Iberian Meseta (PASCAL, 1984; FERNANDEZ MENDIOLA *et al.*, 1989). This emerged zone, a source for coastal clastics, is also lined by platform carbonates in Portugal (Estremadura, Algarve) as well as in the Prebetic Iberian ranges, the pattern of which is controlled by palaeofaults (VILAS *et al.*, 1982). This platform system passes southwards to the Betic basinal sediments (VERA *et al.*, 1982).

III.4.- North Africa

From Egypt to Morocco, coastal clastics fringe the Sahara shield. The Nubian Sandstones Formation from Egypt are typical of this regime and corresponds to "Nubian A", including the *Lingula* shale and the Abu Ballas formation. The corresponding sediments are found in the southern Dakhla basin between Uweinat and Aswan (Kharga uplift) (KLIZSCH & SQUYRES, 1990). The Sidi Aïch sandstones from Tunisia or Sidi R'Gheiss sandstones from east Algeria are equivalent. Near the Tunisia - Libya border, continental clastics are dominant; this facies extends to the Murzuk - Cyrenaïc and Kufra basins in Libya.

A fringing carbonate platform system runs from Egypt to Western Algeria, interrupted in Eastern Algeria by deltaic sediments (Ksour - Sidi R'Gheiss sands, and even the Berthelot sandstones). Carbonates occur widely on the Sahara platform where they are represented by limestones (Serdj Formation) or dolomites, the age of which is either Early or Late Aptian (MASSE, 1984). Platform environments grade northwards to basinal environments. Corresponding sediments are well represented in northern Tunisia (Sillon Tunisien), Tell and North-east Algeria (Sellaoua basin) as well as in North-east Morocco, and correspond with marls and marly limestones including sandy beds, with ammonites (*Deshayesites*). Flysch sediments are known in the Tellian Trough (e.g., El Melaab formation) from northern Tunisia to Morocco (Pre-Rif). They are fed by the Alboran - Kabylia - Calabria bloc and the Sahara-derived clastics (DURAND-DELGA, 1955; RAOULT, 1975; BOUILLIN, 1986). The Constantine Carbonate platform is regarded as an easterly extension of the Hodna platform and is surrounded by slope and basinal environments.

There is a sharp contrast in carbonate platform development between the Atlantic and Mediterranean margins of North Africa, shallow water limestones are nearly absent in Morocco. In the Essaouira - Agadir basins, clastic sediments are dominant, whereas sandy dolomites are locally present. The predominance of clastics over carbonates is well illustrated by the huge deltas located offshore the Tarfaya - El Aïoun region: Tan Tan and West Saharan Deltas (BEHRENS & SIELH, 1982).

III.5.- Syria - Levant

Kurnub continental sandstones are found from the Negev to the Jordan-Israel border, including some volcanism, they grade rapidly westward to sandy then pure carbonates, grading laterally (in offshore settings according to the present Mediterranean coastline) to marls and shales, considered as hydrocarbon source rocks (FLEXER *et al.*, 1986). This transition is relatively short (less than 20 km) and controlled by synsedimentary faults. In the Palmyrides, Early Aptian beds are represented also by continental sandstones (Rutbah sandstones) (PONIKAROV *et al.*, 1966). As in Israel, the sandstones grade north-eastward to carbonates or mixed carbonate-siliciclastics, this limestone unit corresponds to the classical "Falaise de Blanche" which is known from Lebanon to Syria (SAINT-MARC, 1970; MOUTY & SAINT-MARC, 1982).

III.6.- Arabian plate

The Arabian plate shows two kinds of margins (FONTAINE, 1981):

- in the north, from Baïr - Bassit to NE Iraq, shallow clastics and platform limestones grade to a rifted slope (Hezan - Karadut, Kepir) flanking the oceanic domain (Koçali complex);
- in the E-NE from Iraq to Oman, the neritic domain (Zagros, Oman) is flanked basinward by a "radiolarite belt" (Quiqua, Kermanshah, Pichakun, Hawasina), lined by the Bisitoun - Neyriz ridge located at the Ocean/ continent boundary.

At the northern edge of the Arabian promontory, the Mardin Uplift is exposed and surrounded by coastal clastics; similar sediments are recorded from western Iraq and Jordan.

A belt of shallow water limestones is recognised from southern Iraq to Oman, extending to Dhofar. This belt corresponds to the Shuaiba platform (ALSHARHAN & NAIRN, 1986), well exposed in the Oman Mountains. Similar limestones are found in the Zagros Mountains (RICOU, 1976). From the southern part of the Arabo-Persian Gulf down to the foothills of Jebel Akhdar, the Bab basin forms an embayment probably connected to the Tethyan domain through the Musandam region.

Limestone-shale units of basinal significance are reported from northern Oman (Jebel Akhdar), whereas slope to basin sediments with platform-derived bioclastic components occur in the Hamrat Duru Group from the Hawasina nappes (BÉCHENNEC, 1988). In the Pichakun, nappes bioclastic turbidites with orbitolinids document the occurrence of vanished isolated carbonate platforms in the corresponding basin (RICOU, 1976).

IV.- MAIN LITHOFACIES

Four main lithofacies are of special interest during the Early Aptian.

IV.1.- Urgonian limestones

"Urgonian" was first defined as a stage from Orgon (SE France) and remains in use to describe Early Cretaceous limestones (including the Early Aptian) with rudists, corals, dasyclads, benthic foraminifera (especially orbitolinids), bryozoa and stromatoporids, which have a high potential for *in situ* carbonate production. Urgonian facies are a characteristic feature of the Mesogea.

The requienid facies (*Requienia, Toucasia*), the most prominent lithotype among Urgonian facies, corresponds to packstones-wackestones or fine pelleted grainstones with miliolids, whereas caprinid facies (*Offneria, Praecaprina, Pachytraga*) mainly consist of more granular, coarser grained sediments. The orbitolinid facies (dominated by the genus *Palorbitolina*) is a typical member of the Urgonian facies suite. Coral facies, although less common, are also well represented, and may grade laterally or vertically to oolitic/bioclastic cross-bedded grainstones with dictyoconids, dasyclads, non-rudist bivalves, gastropods and echinoderm fragments. Algae-laminated and porostromatic sediments (*Bacinella* dominated) are also common.

The above-mentioned sediment types are made exclusively of pure carbonates and can mix or interbed with siliciclastic or evaporitic deposits to form Urgonian facies complexes.

IV.2- Pelagic limestones

Pelagic limestones are represented by the "Maiolica Facies": micritic carbonates built up mainly by *Nannoconus*, with some coccoliths, calcispheres and radiolarians. Cherts (derived from the dissolution of radiolaria and sponges spicules) are common. The Maiolica was deposited above the CCD, though evidence of deposition below the ACD remains unclear. Depositional depths have been proposed for the latest Jurassic - earliest Cretaceous Maiolica facies from different basinal settings (Central Atlantic, Ligurian basin), mainly based on the SCLATER *et al.* curve (1977). Depth range is from 1000 to 4500 m, with average near 3000 m (WIECZOREK, 1988). In many regions, the calcareous pelagites are just the perpetuation of the Neocomian ones, whereas marly intercalations are frequently more numerous, a tendency which will increase during the Late Aptian, dominated by clays. Therefore, the Early Aptian Maiolica is somehow transitional between a pure calcareous pelagic regime and a terrigenous one, prevailing in the Late Aptian - Albian.

IV.3.- Pelagic marls - carbonate rhythms

Carbonate-marl alternations are a typical feature of basinal environments with a low or regular input of clastic sediment from continental sources. Cephalopods (ammonites and belemnites) are the prominent macrofossils whereas some deep water infaunal and epifaunal bivalves are also present (*Aetostreon, Plicatula, Camponectes*). The microfaunal content is dominated by planktonic foraminifers (*Praehedbergella*) or even deep-water benthic species (*Lenticulina, Epistomina, Praedorothia, Spirillina*). Nannofossils are also well represented. The rhythmic character depends on variations in the production of biogenic carbonates (*Nannoconus*, coccoliths) in the photic zone, relative to the input of the non-carbonate fraction. Rhythmicity is well expressed by the terrigenous content, nannofossil composition (marls contain coccoliths whereas limestones are mainly formed of nannoconids) and ratios in $\delta^{18}O$ and $\delta^{13}C$ (COTILLON *et al.*, 1980). The rhythmic aspect can be modified by diagenesis, especially for sediments deposited near the carbonate compensation depth, where the dissolution of carbonates can occur and alter the original bedding.

Astronomically induced changes of climate and oceanography (in the Milankovitch band) can produce this kind of rhythmic succession. Carbonate is ascribed to high fertility phases correlated with active oceanic circulation driven by relatively high pole to equator thermal gradients (i.e. the "cold" phase), whereas marls with an organic-rich content are ascribed to low fertility phases, correlated with sluggish oceanic circulation driven by relatively low pole to equator thermal gradients (i.e. the "warm" phase) (DE BOER, 1983; HERBERT & FISCHER, 1986).

IV.4.- Black shales

Organic carbon rich sediments (> 1% TOC) are a prominent feature of the Aptian - Albian interval corresponding with the OAE 1 (JENKYNS, 1980). Their depositional depths are similar to those from the pelagic limestones (see above) whereas they are known to have been deposited below or above the CCD, a situation expressed by their variable carbonate content. Relatively "shallow" black shales are also reported from epicontinental settings. PRATT & KING (1986) *in* ARTHUR *et al.*

V.6.- Basin environments

"Shallow" basin environments, i.e. above the carbonate compensation depth, are characterised by fine clastic sediments, pelago-detritic, or nannoplankton-derived carbonates including small amounts of clay, biosiliceous or organic-rich admixtures. Marl-limestone alternations are typical features of the Tethyan basins and are *inferred to reflect cyclic variations in oceanographic* conditions controlled by climatic oscillations.

"Deep" basins, i.e. below the carbonate compensation depth, show carbonate-free sediments and are therefore environments for the deposition of clay and silts or biosiliceous (radiolarites) and organic-rich facies (black shales).

Pelagic limestones (Maiolica facies) are well represented in the Mediterranean Seuil, especially in the Peri-Adriatic domain, where siliciclastic input is very limited.

Marl-limestone alternations are also widespread and recorded in the Atlantic and Peri-Alpine basins when siliciclastic fluxes derived from the adjacent continents are active.

Black shales are known to occupy entirely the Early Aptian interval (central Atlantic), or may occur only as local intercalations in predominantly calcareous (Maiolica-type) or marly successions (eastern Carpathians).

Siliceous radiolarian oozes are known to form in fertile waters (oceanic divergence and upwelling areas) at low latitudes below the carbonate compensation depth (BERGER, 1974). Basinal radiolarites are reported from Oman (Hawasina basin).

V.7.- Hiatuses

The Early Aptian sedimentary record is often missing in many basinal areas.

Two kinds of settings are to be distinguished:

- passive margins with tectonic activity where non-deposition is controlled by prominent topographies favouring condensation or current sweeping phenomena,
- basin plains where deep-water circulation could be involved in preventing sedimentation.

The passive margins of the Central Atlantic and the Western Alps are characterised by fault-rotated blocks on the edges of which sedimentation did not occur. Such tilted blocks of stretched continental crust are well illustrated from the Iberian and West African Atlantic margins.

basin setting hiatuses are reported from the Central Atlantic where the gaps coincide with the change from pelagic carbonate sedimentation to clay dominated deposition. Long duration gaps (including the pre- and post-Aptian time span) are known from the Hawasina basin.

V.8.- Volcanics arcs

Whereas volcano-sedimentary environments are well represented on the Asiatic and on peri-Pacific margins, they are near absent in Peri-Tethyan Mediterranean regions. The Pontides - Armenia volcanic arc belt post-dates the Aptian. Mecksec submarine volcanoes (Hungary) may illustrate this kind of environment, whether this volcanism is linked to rifting or island arc type is not clear (HAAS *et al.*, 1990).

V.9.- Seamounts

Whereas seamounts are a prominent Early Aptian feature in Western Pacific, they are poorly represented in the Mediterranean regions. The Eratosthene seamount is the only example known hitherto from these regions. For MART & ROBERTSON (1998), this seamount pertains to the continental margin of Africa, flanking the oceanic crust of the Levant basin, now subducted northwards along a plate boundary located between the seamount and Cyprus. It consists of shallow water, platform carbonates of early Mid Cretaceous age, capped by Late Cretaceous chalks (borehole 967E, ODP leg 160).

14.- LATE CENOMANIAN (94.7 - 93.5 Ma)

J. PHILIP[1] & M. FLOQUET[1]

I.- MAIN FEATURES

In the Tethyan realm, the Cenomanian stage corresponded to a period of sea level rise that resulted in the Cenomanian transgression. Consequently, Cenomanian deposits are well developed, especially on the cratonic areas that bordered the Tethys Ocean. The maximum transgressive impulse having occurred during Late Cenomanian times, we have therefore selected this substage for mapping (Fig. 14.1).

Cenomanian deposits yield very characteristic fauna (both in platform and basinal facies), which can be easily identified. In the northern Tethys domain (e.g., Eurasia), Late Cenomanian times are subdivided in three ammonite zones, from the base to the top: the Naviculare - Guerangeri Zone, the Geslinianum Zone and the Juddii Zone (WIEDMANN *et al.*, 1978; TRÖGER & KENNEDY, 1996). In southern Tethys (Africa) *Neolobites vibrayeanus*, *Nigericeras gadeni*, *Pseudaspidoceras pseudonodosoides* and *Vascoceras cauvini* characterise the upper part of the stage. In most areas of marginal basins, planktonic foraminifera are useful for biostratigraphic correlation. The zonal scheme used here is based on previous studies (ROBASZINSKI & CARON, 1979, 1995; CARON, 1985). The Cenomanian - Turonian transition is marked by the extinction of *Rotalipora cushmani* and by the occurrence of *Whiteinella archaeocretacea*. *Praeglobotruncana helvetica* appeared in Early Turonian times.

In basinal facies, reliable data may be provided by additional fossil groups such as deep-water benthic foraminifera (KUHNT & MOULLADE, 1991), calcareous nannoplankton (BRALOWER, 1988), radiolaria (THUROW, 1988), dinoflagellates (HERBIN *et al.*, 1987), and inoceramids. In the carbonate platform facies, extensively represented during Cenomanian times, rudists (PHILIP, 1978) and large foraminifera (NEUMANN & SCHROEDER, 1981) provide accurate records for dating.

According to HAQ *et al.* (1987) and ODIN (1994), the duration of the whole Cenomanian is 4 Ma, from 96 to 92 Ma. However, GRADSTEIN *et al.* (1995) and HARDENBOL *et al.* (1998) estimate the duration of the Cenomanian times to 5.4 Ma, comprised between 98.9 (+/-0.6) and 93.5 (+/-0.2) Ma.

Accurate geochronologic datings of the Late Cenomanian ammonite Zones have been recently established by HARDENBOL *et al.* (1998). These authors have placed the base of the Guerangeri - Naviculare Zone at 94.71 Ma and the top of the Juddii Zone at 93.5 Ma.

Since the Cenomanian is included within the long C 34N magnetic Chron / Polarity Zone, some imprecisions remain in the establishment of the palaeogeodynamic frame, especially about the palaeoposition of moving blocks.

Sedimentological and palaeontological data indicate a warm, equable climate for the Tethyan realm during Cenomanian times (FRAKES & FRANCIS, 1990).

II.- STRUCTURAL SETTING AND KINEMATICS

Positions of Laurasia and Africa megaplates have not been deduced from palaeomagnetic data due to the fact that the Cenomanian stage belongs to the Cretaceous quiet C 34N.

A set of intraoceanic thrust planes east of Arabia corresponded to the initial step of the peri-Arabic obduction. Its location is deduced from the position of oceanic ridges before obduction.

The ridges that limited the transit space from Africa have been reconstructed from the peri-Arabic ophiolites, now obducted. The corresponding directions of spreading axes for the Taurus and Oman parts require the existence of a third branch. This branch separated an Apulian plate from a Transit plate (off-India) and was somehow inherited from a Late Jurassic major transfer fault. Its location has been chosen to account for the Cretaceous break-up and ophiolite formation in Iran.

The position of the palaeomagnetic pole is taken from BESSE & COURTILLOT (1991).

III.- DEFINITION OF DOMAINS

III.1.- West Siberian

Mostly covered by clastic terrigenous deposits of open marine shelf, this domain extended NE of the today

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Ural Range. Depositional environments were clearly polarised: proximal to the west along the exposed Uralian areas, which provided the terrigenous facies, distal to the east where hemipelagic conditions developed. This domain possibly communicated south-eastward with the Mangyshlak area through the Turgay passage.

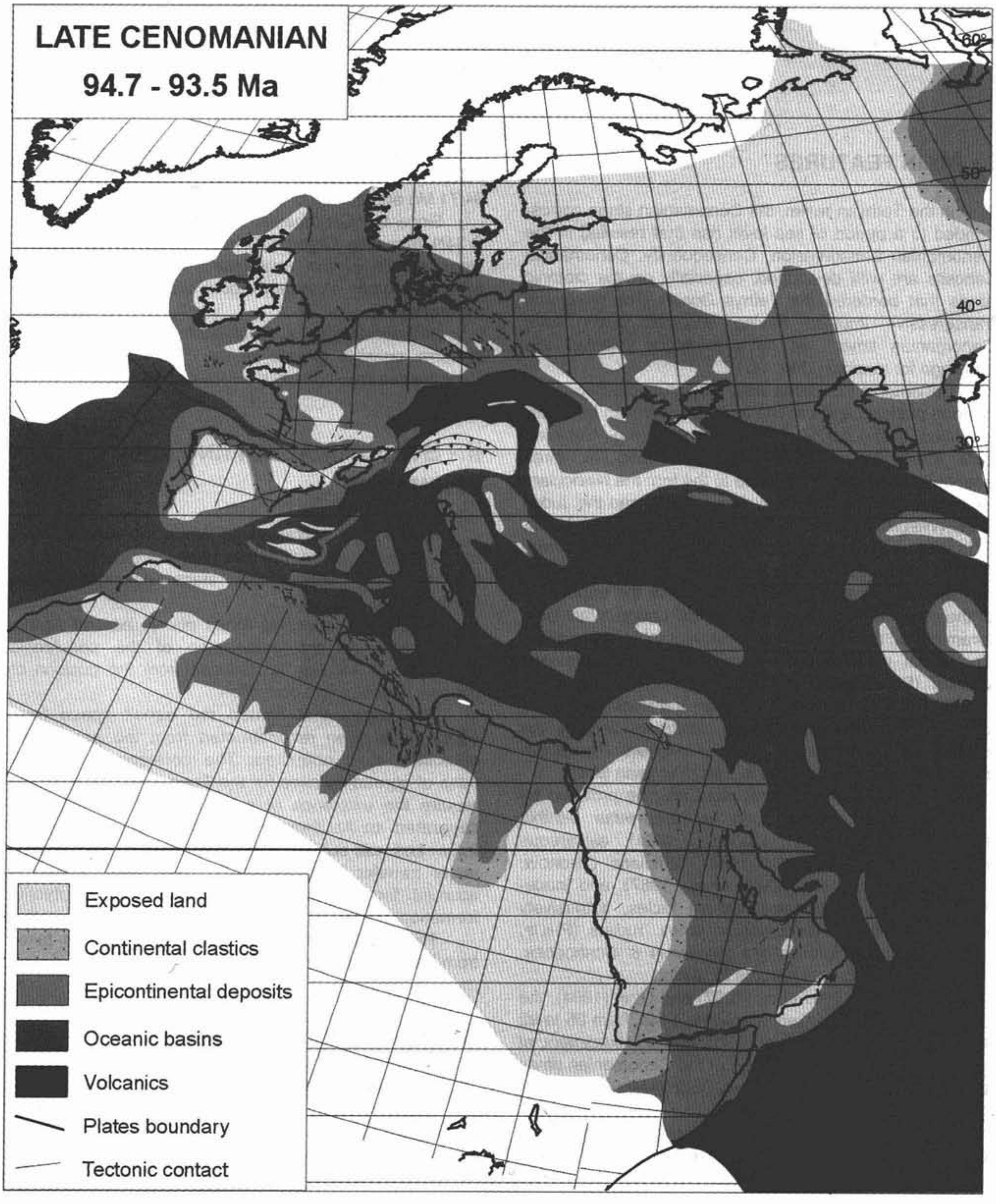

Fig. 14.1: Simplified palaeogeographic map of Peri-Tethyan area during the Late Cenomanian.

III.2.- Russian platform, Scythian platform, Mangyshlak, Turan

During Late Cenomanian times the whole of these areas functioned as a same palaeogeographic domain. The assumed exposed lands corresponding to the Russian craton and to the present southern Ural Range were the sources of the clastic facies deposited in the coastal marine environments.

In the central areas, sandy facies spreaded out in a very shallow marine water belt, of several hundreds kilometres wide, which formed the main part of the Russian platform. From Moscow, on the north, to Kupyansk, on the south, only calcareous sands, sometimes with phosphatic nodules horizons, occurred. Dating is based mainly on the presence of *Preactinocamax plenus* belemnite, *Lingulogavellina globosa* benthic foraminifera and *Microrhabdulus decoratus* calcareous nannofossil (ALEKSEEV, *in litteris*, 2000). Southward, between the present Azov Sea and Caspian Sea areas, hemipelagic conditions developed in a wide-open shelf directly connected to the deep Tethyan basin. The deeper part of this basin was the Great Caucasus trough where mainly flysch sedimentation occurred (NIKISHIN *et al.*, 1998a).

In the eastern areas, in the western part of the today Precaspian depression, Turonian deposits directly overlie Early Cenomanian sands, without sediments including fauna of Late Cenomanian age. To the south, calcareous sands with *Praeactinocamax plenus* and numerous planktonic foraminifera accumulated. Farther southwards, in Turan, facies graded to clayed limestones and marls indicating open shelf depositional environments.

In the western areas, the Ukrainian Crystalline shield and the today Donetz basin were exposed low lands, sources of terrigenous facies. To the north-east of the Ukrainian shield sandy marls accumulated in the Dniepr - Donetz depression while northward chalky and marly facies deposited in the Ukrainian basin. To the south and south-east (on the northern margin of the Black Sea) sandy limestones and marls characterised conditions of hemipelagic and open shelf (Scythian outer platform).

In many places, condensed chalky deposits including gaps and phosphoritic horizons characterise the Late Cenomanian and the Cenomanian - Turonian boundary (ILYIN, 1994; BARABOSHKIN *et al.*, 1998). Such facts probably express a major deepening and transgressive event at that time.

III.3.- Moesian platform

During Cenomanian times, this domain was bounded southwards by the Rhodope massif. The Moesian platform corresponded to a hinge between the Alpine trough to the west, and the Black Sea basin to the east. This domain was likely affected by a post - Aptian to pre-Maastrichtian dominantly extensional phase (BERGERAT *et al.*, 1998).

Shallow marine coastal deposits, and possibly in some places continental terrigenous facies, dominated during Late Cenomanian times in the area between North Dobrogea and Balkan thrust belt. Carbonate shelf deposition occurred in the east and north-west of Moesia and in pre-Carpathian zone where carbonates of deeper open shelf accumulated also (GEORGIEV, *in litteris*, 2000).

In South Dobrogea, the Late Cenomanian series correspond to a 60 m sequence of grey marls and grey sandy marls rich in planktonic foraminifera (AVRAM *et al.*, 1995-1996).

III.4.- Crimea

The main peculiarity of the Cenomanian deposits of Crimea is the rhythmically bedded succession of marls and limestones, indicative of deep environments of the outer part of the shelf. The Cenomanian - Turonian boundary is well characterised by anoxic organic - rich sediments (black shales beds) with silt-size quartz grains, glauconite, volcanoclasts and lack of benthic fauna.

III.5.- Black Sea

Major post-rift subsidence and probable oceanic crust emplacement affected the Western Black Sea during Cenomanian times (SPADINI *et al.*, 1996). Cenomanian deposits unconformably overlie the syn-rift Aptian to Albian sediments. They consist in a unit of pelagic carbonates and distal tuffs that are interpreted as proofs of the change from rifting to drifting in the western Black Sea (GÖRÜR *et al.*, 1993). The Western Black Sea developed off the stable continental Moesian platform in a setting generally considered to be of "back-arc" type, but without any contemporaneous volcanics. The lithosphere in this case would be expected to be both thick and cold (SPADINI *et al.*,1996). According to these authors, the eastern Black Sea possibly settled to the north of major volcanic arcs, now exposed in the eastern Pontides. Thus, it can be considered to have formed in a back-arc setting.

III.6.- Polish Trough

The Mid-Polish Trough developed during Mesozoic times over a deep and important structure known as the Trans European Suture Zone (TESZ) between the East European Craton and the West European platform. In the Polish area the TESZ coincides with the Teisseyre - Tornquist Zone (TTZ) being an old crustal zone (SWIDROWSKA & HACKENBERG, 2000).

During Cenomanian times, the trough enlarged significantly (MAREK, 1997). It widely communicated with the Ukrainian basin and farther east with the Russian platform. The connection with the Tethyan domain opened again.

III.7.- North European basins

This area was bounded northward by the Scandinavian cratonic area and southward by the Sudètes, Bohemian and Rhenish massifs, and included some isolated basins such as the Danish trough, the Sub-Hercynian basin and the Munster basin. The Pompeckj block is characterised by pelagic calcareous sediments

consisting of limestones, chalk and opoka (siliceous chalks) (NIEBUHR & PROKOPH, 1997).

Around the probably emerged southern massifs, the Cenomanian clastic deposits lie transgressively on the Albian series and, in many areas (as the Munster basin) directly cover the Palaeozoic substrate (MARCINOWSKI, 1974).

The Bohemian Cretaceous basin is located in the central part of the Bohemian massif. Although it is nowadays separated from the neighbouring Cretaceous basins, palaeogeographic reconstructions show that there was communication between them through narrow straits during a period of high sea level (VALECKA & SKOCEK, 1991).

The most conspicuous sea level change is associated with the Cenomanian - Turonian boundary. Cenomanian sandstones and siltstones are overlain by Early Turonian marlstones or limestones, which are transgressive in marginal parts of the basin and onto intrabasinal elevations of pre-Cretaceous basement. ULICNY *et al.* (1997) have established the relationships between geochemical anomalies, bioevents, and sea level rise at that time.

On the Scania and Swedish coast, Cenomanian deposits (sands and limestones) rest upon the crystalline basement. This area was entered by the transgression during Middle - Late Cenomanian times.

III.8.- British Islands and North Sea

In the North Sea area, quieter fully marine conditions of regional subsidence, centred over the axial graben system, prevailed during Late Cretaceous times. The stable massifs and highs were passive except for some faulting marginal to the Viking graben and, to a much lesser degree, in the Central graben. Almost all the North Sea was basinal in a facies sense. Most of the Late Cretaceous deposits of the southern North Sea are chalk. North of latitude 57°, a clastic component enters and chalk decreases in proportion (HANCOCK, 1990). Cenomanian glauconitic sandstones occur near the borders of the emerged lands (Grampian High, Irish massif, Welsh massif, Cornubia). The latest Cenomanian formation (Plenus Marls) is commonly composed of grey or black, variably calcareous soft claystones, presenting a high organic content in the darker layers.

III.9.- Western European

A SE-NW orientated high acted as a major boundary between the north-eastern European chalky domain and the south-western European shallow carbonate domain. This high comprised the assumed emerged Armorican massif, the Massif Central plus its south-eastward extension, the Durancian swell (DU). Wide coastal belts mainly constituted of clastic deposits surrounded these areas.

To the north and to the north-east of this high, quite pure chalk, very rich in coccoliths, deposited in the Anglo-Paris basin, indicating open sea conditions due to wide connections both toward the North Sea and toward the Alpine basin. Ammonites bearing marls and clayed limestones deposited in this basin.

To the south of the high and its terrigenous edge, carbonate platforms developed, such as the Aquitanian (Aq), which was open towards the Central Atlantic, as well as the North Pyrenean and Provençal ones.

The WNW-ESE elongated Cantabrian - Pyrenean - Provençal trough corresponded roughly to the boundary between the European plate and the Iberian microplate. Deep basinal marly facies accumulated in this trough.

The Iberian microplate supported two main lands: the Iberian massif and the Ebro massif to which the Corso-Sardinian massif (CSb) was linked. A seaway, the Iberian Strait, crossed over the microplate and made possible connections between the Atlantic and Tethyan oceans.

Carbonate ramps extended all around the massifs as in southern Pyrenees, in the Castilian, Lusitanian (LB) and Valencian areas.

Due to marine transgression and deepening during Late Cenomanian times (Naviculare Zone), carbonate ramps generally were homoclinal and under open sea conditions, which explains why rudist build-ups did not thrive as well as on the previous Middle Cenomanian well zoned platforms.

III.10.- Maghreb

This domain was characterised by a broad extent of shallow carbonate platform facies, as a result of the Late Cenomanian transgressive event.

In Morocco, the transgression reached its maximum and the shelf facies became homogeneous within the whole Atlasic domain. To the west, they covered the Western Meseta. To the north, the Idrissides High (CHOUBERT & FAURE-MURET, 1962) constituted the northern border of the seaway. Late Cenomanian rudist rich carbonate facies deposited to the south in the Anti-Atlas area (FERRANDINI *et al.*, 1985), grading to the north in the Middle Atlas area to outer shelf carbonates with ammonites (CHARRIÈRE *et al.*, 1998).

In southern Algeria, the latest Cenomanian carbonate platform progressively changed into shallow cephalopod rich carbonates, but with very scarce benthic organisms (BUSSON, 1972; BUSSON *et al.*, 1999). Disturbed palaeoceanographic conditions (dysaerobic) (Fig. 14.2) have been inferred in order to explain this faunal impoverishment. The northern Algerian platform graded northward to basinal environments (Aurès range).

In Tunisia, the Cenomanian sequence exhibits clear syndepositional features along NNW-SSE trending normal faults. They gave rise to tilted blocks above décollement layers, intraformational breccias and slumps (BOUAZIZ *et al.*, 1998). In southern Tunisia (Gafsa and Chotts area), the onset of the carbonate sedimentation took place during Late Cenomanian times as a result of the general transgressive impulse (RAZGALLAH *et al.*, 1994). The Cenomanian - Turonian transition is marked by an anoxic event (ABDALLAH & MEISTER, 1996) followed by a shallowing upward of the carbonate platform and a northward progradation of rudist rich facies (RAZGALLAH *et al.*, 1994). The carbonate platform evolved northwards to the North Tunisian basin. In this basin, a continuous argillaceous - marly deposition prevailed, while events (storms?) occurring in the platform produced episodic

clastic sediment supplies. During latest Cenomanian times, the Bahloul laminated anoxic formation (Fig. 14.2) deposited (ROBASZYNSKI *et al.*, 1993). In north-eastern Tunisia, an isolated carbonate platform settled, probably linked westward to the Constantine platform. Carbonate slope deposits, including mass flows with reefal components, originated from the south margin of this north-eastern Tunisian platform.

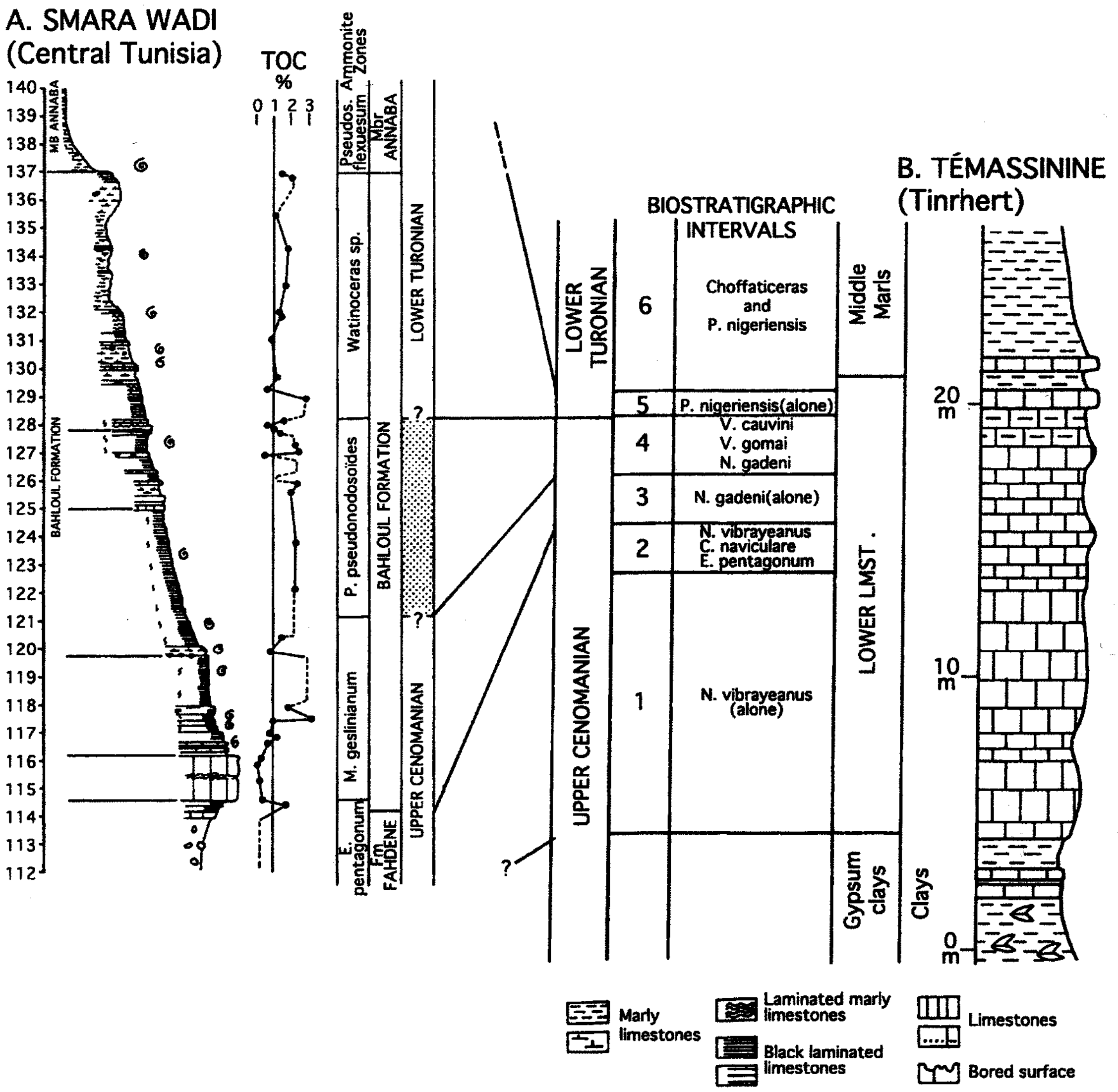

Fig. 14.2: Cross section of Late Cenomanian to Early Turonian series at Smara Wadi, near Kalaat Senan, in Central Tunisia (from ACCARIE *et al.*, 1996) and at Témassinine (Tinrhert) showing stratigraphical relationships and main facies in both domains, which are separated by more than 800 km (from BUSSON *et al.*, 1999).

III.11.- Sirt basin

A synthetic overview of the Late Cretaceous series has been presented by WENNEKERS *et al.* (1996). The existence of an irregular Late Cretaceous palaeotopography, together with contemporaneous structural activity, controlled the distribution, facies and thickness of the sediments.

Evaporitic sedimentation took place during Early Cenomanian times. It was followed by uplift and erosion during Middle Cenomanian times. This resulted in an eroded landscape broken into irregular, generally NW trending horst and graben structures, which were gradually submerged during the subsequent major Late Cenomanian marine transgression. Sandy carbonates (Lidam formation), showing strong variations in thickness, can be assigned to Late Cenomanian times.

III.12.- Egypt - Sinai Peninsula

A stratigraphic synthesis of the series of this area was recently presented by KUSS & BACHMANN (1996). The area of NE Egypt, Sinai and South Israel was situated between two major regional tectonic units: the stable Arabo-Nubian shield, to the south, and the Alpine mobile orogenic belt, to the north.

The southward onlapping marine Cretaceous succession overlies lithologies of different Phanerozoic ages. Thus, it unconformably covers continental-fluvial sandstones of Jurassic to Early Cretaceous age in the North. Further to the south, the Cretaceous strata rest disconformably upon the Triassic - Cambrian or Precambrian basement rocks.

The distribution of extensive shallow subtidal/ peritidal Middle - Late Cenomanian carbonates documents a maximum flooding of the sea, followed by a gradual northwards retreat during Early Turonian times (KUSS & BACHMANN, 1996).

During Cenomanian times, the ongoing sea-level rise led to the Tethyan wide sea level highstand (PHILIP *et al.*, 1993). As a consequence, Cenomanian strata cover broad areas. They are made of marine siliciclastics, mixed with carbonates or interfingering with pure limestones. A conspicuous regional feature was a gulf-like embayment located at the present Eastern Desert. The Late Cenomanian shelf facies of this Eastern Desert and of northern Sinai is characterised by a rich assemblage of benthic foraminifera, among them *Praealveolina cretacea, Pseudedomia drorimensis* and also ammonites of the genera *Neolobites*, *Vascoceras* and *Acanthoceras* (KUSS & MALCHUS, 1989). These ammonites were also described in southern Israel (LEWY & RAAB, 1976).

Cretaceous palaeostress evolution in north-western Egypt has been established by MOSTOFA (1999). During Late Cenomanian times, the tectonic evolution of this area started by a NW-SE trending extensional phase of deformation dominated by NE-SW trending synsedimentary fault populations.

III.13.- Syria - Levant

Main parts of the Levantine and Syrian domain were opened to the west, towards the eastern Mediterranean oceanic basin, during Late Cenomanian times. Thus, the marine transgression during the Naviculare Zone led to a clear W-E environmental polarity.

Open marine and distal environments dominated westwards while proximal coastal environments onlapped eastward the Rutbah High, northernmost promontory of the Arabian shield. Ammonite (especially *Eucalycoceras pentagonum*) bearing marls and argillaceous limestones covered almost the entire Levantine and Syrian domain at the time of the maximum flooding. This flooding is particularly well recorded in the sedimentary series cropping out in the Coastal Ranges and in the South-western Palmyrides. Coastal terrigenous environments retreated to the east and the south on the Arabian shield.

Shallow carbonate ramps settled later as the sea level rise ceased or slowed down. Rudist build-ups (including Caprinulids and Radiolitids) formed and yielded coarse bioclastic calcarenites that accumulated in large clinoforms located on upthrown fault blocks. Finally, most of the carbonate ramps probably became subaerially exposed at the end of the Cenomanian times. Calcarenites often appear porous and very similar to the ones of the Mishrif Formation.

The Euphrates graben (EuG) appeared in Central and Eastern Syria. Late Cenomanian times corresponded to the transition period between an early rifting phase (N 120 - N 140 extension and associated volcanic activity) and an active syn-rift phase (N 10 extension and development of the carbonate sedimentation) (JAMAL, 1998).

III.14.- Arabian platform

In the Arabian Peninsula (Iraq, United Arab Emirates, Oman), the shelf area became increasingly carbonate-dominated during Cenomanian times, when deposits of the Rumaila and Mishrif Formations of the Wasia Group onlapped the Arabian craton. Carbonate sands deposited in association with build-ups of caprinid and radiolitid rudists that either rimmed intrashelf basins (PHILIP *et al.*, 1995; VAN BUCHEM *et al.*, 1996) or developed on upthrown fault blocks. Submarine fans of shell debris were distributed over the adjacent shelf and basin (HARRIS & FROST, 1984).

According to WATTS & BLOWE (1990), major changes in the depositional patterns on the Arabian carbonate platform succession occurred during Cenomanian times in response to tectonism along the Oman (Oma) margin, and the early development of the Aruma Intrashelf basin. Locally, porous calcarenites of the Late Cenomanian upper part of the Mishrif or Natih Formation in Oman formed on tilted upthrown fault blocks. Major changes are also recorded on the Cenomanian slope sequence, where the transition from carbonate to siliceous facies might be related to tectonism and reduced sedimentation of peri-platform carbonates along the Oman continental margin.

The Wasia - Aruma break within the platform sequence is a major unconformity separating the Late Cenomanian series or older formations of the Wasia Group and the Coniacian series or younger formations of the Aruma Group. Emergence of the platform and development of the unconformity may be related to tectonic activity along the Oman margin and migration of a peripheral bulge over the outer Arabian platform (ROBERTSON, 1987).

Towards the Arabian shield, the Cenomanian carbonate platform was affected by clastic and clay inputs.

III.15.- Alpine Tethyan domain

Palaeoenvironmental and palaeogeographic units of this domain have been reported according to the previous Late Cenomanian map (PHILIP *et al.*, 1993). Some revisions have been carried out by POISSON (*in litteris*, 1999), for the Anatolian platform and the surrounding basinal areas.

In the northern Neo-Tethys, the mid-oceanic ridge ceased to be active in some places. At that time, plankton-bearing chalks covered some of the ophiolitic melanges that formed previously in deep basins. Carbonate platforms (i.e. Kirsehir) existed at the same time producing detritals, which were resedimented (debris flows) in surrounding basins. Some continental rises were eroded and flysch-like deposits invaded other small basins.

The Anatolian platform was the site of shallow marine carbonate deposition, while some areas emerged, thus allowing bauxite formation.

The Pamphylian basin was always the site of deep sediment deposition. In some places, radiolarites replaced the carbonate chalks indicating a deepening of the basin. During latest Cenomanian times, the Bey Daglari platform was broken by a system of normal faults and large parts of it collapsed in deep basinal conditions. Coarse breccias formed along the margins of the adjacent Pamphylian basin, as a result of a strong erosion of the fault scarps.

Subsequent to the opening of the Mediterranean basin to the south, the Bey Daglari platform migrated northwards while the Eratosthenes area remained approximately at its present day position. The major fault which separated the Bey Daglari and Eratosthenes since Early Cretaceous times, remained active.

IV.- PALAEOENVIRONMENTS AND FACIES

IV.1.- Exposed lands

On the northern margin, the Fenno-Scandian shield played an important role, bounding the Cenomanian marine shelves and providing the major part of clastic quartz and clay components. Farther on the east, the Ural Range area acted as a palaeogeographic barrier separating the West Siberian domain from the Russian craton and platform.

From the Ukrainian area to the east, up to the Western European domain to the west, isolated exposed massifs extended, surrounded by clastic or carbonate shelves. Of low relief, these massifs (e.g., Ukrainian shield, Bohemian and Rhenish massifs) have provided small amounts of clastic materials. It seems that they also have been largely flooded by the Late Cenomanian transgression. Furthermore, the Western European massifs formed a palaeogeographic barrier, which isolated the European shelves from the Central Atlantic Ocean.

On the southern margin, the Saharan and Arabian shields were a broad continental exposed area, providing fine mainly clayed inputs in the Late Cenomanian sea. The shields were interrupted by seaways, such as in Tibesti and Tademait, which linked the Saharan platform to the South Atlantic.

IV.2.- Fluviatile, lacustrine, fluvio-lacustrine

During this time of general flooding, very few areas were the setting of a continental deposition. This type of sedimentation thus occurred in peculiar palaeogeographic conditions such as very-inner parts of gulfs, bays or inlets where continental palaeoenvironments have been preserved.

For example, continental deposits exist at the western border of the West Siberian gulf, around some exposed lands of the West European domain (such as the upper part of the Utrillas formation on each side of the Iberian Strait and as the very upper part of the "Bellasian" sandy facies on the easternmost rim of the Lusitanian basin), south of Libya and along the east side of the Arabian shield.

IV.3.- Coastal marine, shallow marine (terrigenous - clastic)

On the northern margin, a mixed siliciclastic (sands and clays) carbonate - glauconitic sedimentation occurred: 1) on the southern rim of the Fenno-Scandian shield, 2) on the east and south of the today Ural Range area and 3) all around the emerged central and western European massifs not entirely covered by the Late Cenomanian transgression. Palaeoenvironments corresponded to shallow marine conditions with both infaunal and epifaunal associations dominated by molluscs.

On the southern margin, fine siliclastic sedimentation (mainly clays and sands) occurred south of Libya and Egypt, east of Arabia. Shallow environments with fluctuating salinities are sometimes associated with this sedimentation in southern Egypt and maybe on the southern Saharan platform.

IV.4.- Carbonate platforms

The northern and southern Tethyan carbonate platforms differed mainly in climatic and oceanographic conditions.

The northern ones were influenced by wet temperate climatic conditions as proved by the large distribution of coals deposits on the Laurentia and Angara continental areas (PARRISH *et al.*, 1982), although the marine sedimentation was characterised by the input of fine terrigenous sediments, mixed with chalk deposits. North-South cold-temperate currents moving through seaways between the Tethyan and Boreal seas prevented the enlargement of the carbonate platforms and favoured stratification of the oceanic water column and a stagnation of the bottom waters (ARTHUR *et al.*, 1987).

In contrast, the southern Tethyan Late Cenomanian platforms developed in a relatively arid climate, as indicated by some coastal evaporitic deposits. Owing to the warm and shallow water, rudist bearing carbonate platforms extended broadly along the northern margin of the African plate, the eastern margin of the Arabian and the western margin of Eurasia (Aquitaine gulf, Pyrenean areas, Iberian Strait and Lusitanian gulf).

The Cenomanian - Turonian boundary is characterised by a general demise of carbonate platforms and of aragonite secreting rudists (PHILIP & AIRAUD-CRUMIÈRE, 1991). As a consequence, hard grounds and stratigraphic hiatuses occurred at that time in main Tethyan carbonate platform areas. The major demise stage of carbonate platforms is coeval with the Archaeocretacea Biozone, and contemporaneous with the global oceanic anoxic event (OAE_2).

IV.5.- Hemipelagic environments

On the northern Tethyan margin, chalky hemipelagic deposits prevail west of the Ukrainian shield up to the Central Atlantic. The abundance of coccoliths attests to dominant open sea conditions with minor terrigenous sedimentation and favourable oxygenation. At the opposite, east of the Ukrainian shield, hemipelagic deposits are mainly argillaceous limestones, marls and sandy marls rich in planktonic foraminifera.

On the southern margin, the hemipelagic succession is mainly made of alternating marls and limestones rich in ammonites and planktonic foraminifera such as *Rotalipora* and *Hedbergella*.

The peculiar setting of the chalky sedimentation and of the rhythmic marly - calcareous one could be explained by different latitudinal climatic conditions at this time.

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15.- EARLY CAMPANIAN (83 - 80.5 Ma)

J. PHILIP[1] & M. FLOQUET[1]

I.- MAIN FEATURES

The Campanian stage is widely represented on both north and south margins of the Tethys domain. While Late Santonian times were marked by regressive and tectonic global events, Campanian times were characterised by a new broad transgression, enhanced by a relative quiet tectonic activity on the Peri-Tethyan continental platforms. Accordingly, the Campanian deposits are predominantly marine and characterised by a rich content of organisms.

Owing to its well-identified biostratigraphic data, the early part of the stage has been chosen for mapping. It corresponds, for the most part, to the Bidorsatum ammonitic Zone, which is well represented in Europe (HARDENBOL *et al.*, 1998). Belemnites also give accurate data in NW European, Balto-Scandia, and Russian platform chalky domains. The Early Campanian fits in well with the *Globotruncanita elevata* Interval Zone, which ensures precise correlations between north and south Tethyan margins. Early Campanian carbonate platform facies contain generally abundant large foraminifera (*Orbitoides tissoti, Siderolites vidali*) and rudists.

HARDENBOL *et al.* (1998) distinguish three major transgressive-regressive facies cycles for the whole Campanian times. According to these authors, the Santonian - Campanian boundary fits in with the maximum flooding of the first cycle initiated during latest Santonian times.

The duration of the Campanian stage is of 11 Ma for ODIN (1994) and 11.2 Ma for GRADSTEIN *et al.* (1995) and HARDENBOL *et al.* (1998). Taking into account the error margin given for the geochronologic dates (+/-1Ma for ODIN (1994), +/-0.5 Ma for GRADSTEIN *et al.*, 1995), the 83 Ma dating for the lower boundary of the stage seems in good agreement with the different works while, according to GRADSTEIN *et al.*, 1995, the Early - Middle Campanian boundary could be fixed at 80.5 Ma.

Early Campanian times correspond to the C33R magnetic chronozone.

II.- STRUCTURAL SETTING AND KINEMATICS

When it came to draw the Early Campanian Peri-Tethyan areas (Fig. 15.1) of the Laurasian and African megaplates shown in this work, the positions adopted were quite the same as those chosen for the Late Maastrichtian map. Therefore, they are based on the position of the palaeomagnetic pole given by BESSE & COURTILLOT (1991) for Late Maastrichtian times. However, the oceanic space between Eurasia and Africa may have been broader during Early Campanian times, while latest Cretaceous compressive movements just arose.

The major global tectonic event occurred during Late Santonian times, from 85 Ma up to 83.5/83 Ma. It led to a major structural reorganisation that strongly influenced the further successive Campanian and Maastrichtian palaeogeographies. Basically, it was a compressional event directly linked to the change in poles of rotation for the opening of the Atlantic (GUIRAUD & BOSWORTH, 1997). This corresponded to the end of the long and quiet Cretaceous 34 normal magnetic Chron/Polarity zone.

Numerous previous sedimentary basins along the southern Tethyan margin, from Morocco to the Syrian Arc, were folded (GUIRAUD, 1998; GUIRAUD & BOSWORTH, 1997). In Oman, along the north-eastern margin of Arabia, ophiolites began to be obducted (LE MÉTOUR *et al.*, 1995). At the same time, the European Alpine chains suffered converging movements due to the northward drift of the Apulian plate that followed the south-eastward moving of Africa. Subsequently, the Iberian plate drifted north-westward. The Ligurian Ocean was almost subducted (STAMPFLI, 1993) while the Valaisan Ocean and, to the south-west, the Pyrenean trough began to close. Major discontinuities (especially well expressed by breaks in subsidence curves) in the sedimentary series of basinal and platform environments were response to the tectonic activity. In some places these discontinuities are dated around the Santonian - Campanian boundary as in Austria (WAGREICH, 1993), or as in the subalpine basin in south-eastern France (FRIÈS, 1999), or at about 84/83 Ma, as in northern Spain (FLOQUET, 1998).

After this Late Santonian tecto-event, several phases of rifting and downwarping or passive margin

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development occurred during the Campanian - Maastrichtian time span and particularly during Early Campanian times. The acceleration of subsidence led to the rejuvenation of many basins along Peri-Tethyan margins as well as within intraplate areas.

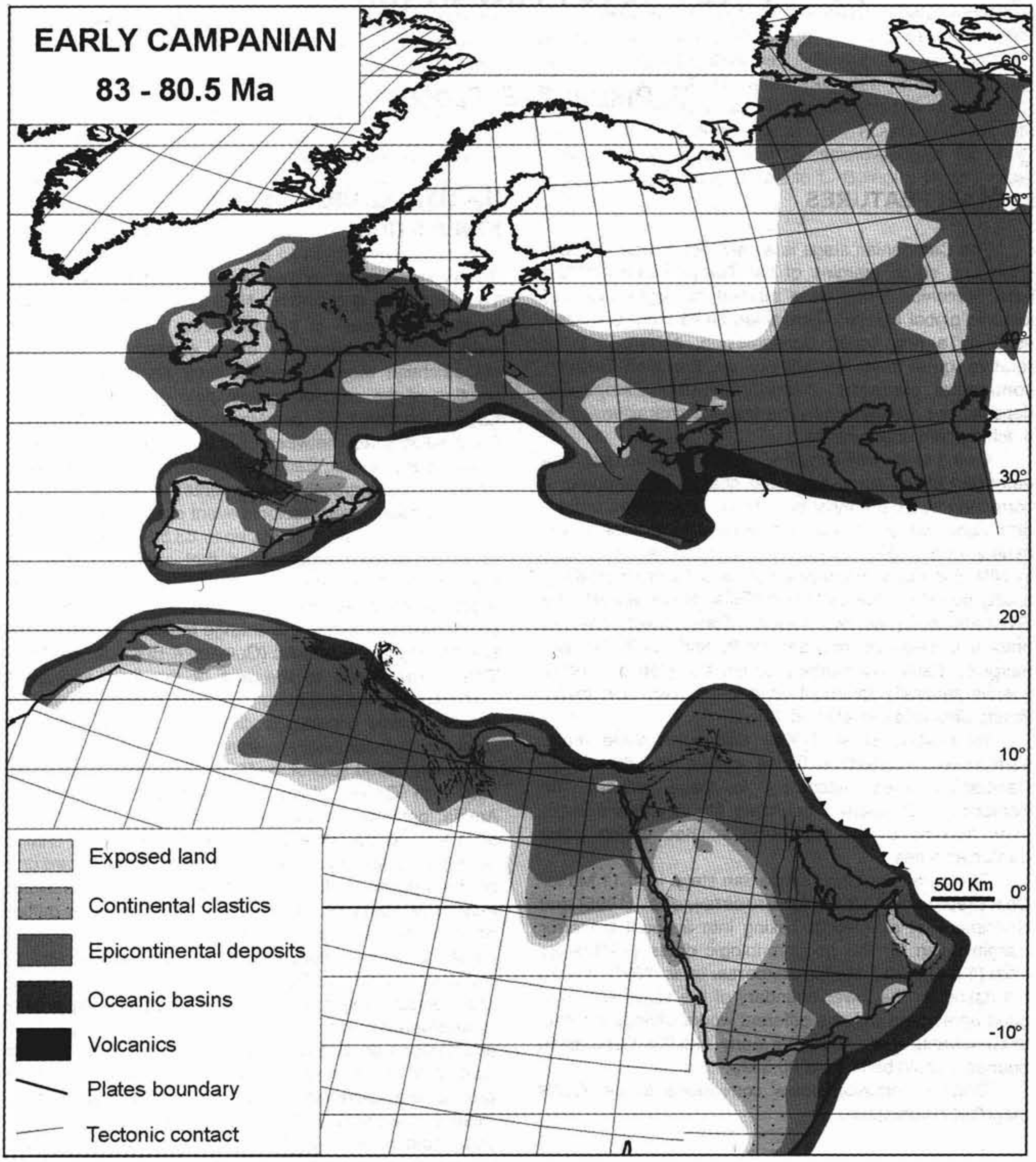

Fig. 15.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Campanian.

III.- Definition of domains

III.1.- Ural

During Late Cretaceous times, the Ural Range country and adjacent regions did not represent an uniform consolidated land. During Turonian - Maastrichtian times, the Polar - Uralian Strait, crossed sublatitudinally the Uralian land and separated the northern Paikhoy - Novozemelsky Range from the more southern north Middle Uralian Range. This strait connected the Pechorian basin with the West Siberian basin.

During Late Santonian - Early Campanian times, the narrow and shallow Middle Uralian Strait brought together the Russian platform and the West Siberian Sea.

Along the whole eastern border of the Ural Range, longshore currents transporting water masses and organism larvae from the south led to the enrichment of the boreal - arctic biota of the restricted West Siberian Sea with Tethyan and Peri-Tethyan species.

Sandy and silty coastal marine sedimentation prevailed in the Turgay area, grading laterally to sandy bearing pollen and spore continental deposits (AMON *et al.*, 1997).

III.2.- Russian platform, Mangyshlak, Turan, Scythian platform, Ukrainian basin

The Campanian succession developed broadly on the Russian platform. It is relatively thick and is formed by marls and chalks with abundant belemnites and benthic foraminifera typical from Western Europe. Campanian times corresponded to the largest marine transgression on the Russian platform during the Late Cretaceous period.

Northeast of the present Azov Sea, a deltaic complex formed, belonging to a large river system which flowed from the north-west close to the modern Don river valley. Around this delta settled a wide belt of siliceous sediments, which were related to high productivity of siliceous organisms (diatoms) enhanced by fresh water supply.

In the Mangyshlak area, a faunal and sediment condensation shows the Santonian - Campanian boundary. The Early Campanian deposits yield abundant belemnite rostra and scarce inoceramids. Deposits are dolomitised marls (KOPAEVICH *et al.*, 1999).

The Turan platform was covered mainly by carbonates grading southward to marls of deeper marine water. These deposits extended to the Precaspian depression and to Crimea areas.

In the Scythian platform, limestones and marls deposition prevailed during the whole of the Campanian times.

In the Ukrainian basin, Early Campanian times, identified by *Gonioteuthis quadrata,* are represented by marls grading south-westward to clayey marls, siltstones, sandstones and phosphatic - glauconitic beds.

The uplifted western part of the Ukrainian Crystalline Shield was source area for sand bodies deposition in Western Ukraine. Emerged areas also existed in the surroundings of the modern Donetz region and were local sources of clastic material.

III.3.- Moesian platform

The Santonian - Early Campanian succession (Murfatlar Formation) is recognised by both its exposure and subsurface in the eastern part of Southern Dobrogea. The main unit, 35 m thick, is built up of bioturbated chalk with cherts, rich in foraminifera and nannofossils, witnessing to offshore shelf conditions (AVRAM *et al.,*1995 - 1996).

Shallow shelf carbonate deposition dominated in Moesia while deeper marly depositional environments existed only in North-Western Moesia and the eastern Black Sea areas.

Terrigenous coastal - shallow marine facies took place in Western Moesia (Lom depression).

III.4.- Polish Trough

Tectonic activity persisted during Campanian times (MAREK, 1997) and was recorded by sedimentation in the Polish Trough. Calcareous and opoka lithofacies were dominant but chalky facies, which accumulated from Early Turonian in the eastern part of the trough, now appeared in its western and southern parts. The amount of terrigenous material within the trough considerably decreased, whereas opokas developed in the outer basinal areas.

The Early Campanian biostratigraphic division has been based on *Gonioteuthis* index belemnites, which were found in the areas where opoka lithofacies dominates.

III.5.- German basin

In North Germany, marginal Late Cretaceous sediments are represented by two vertically stacked units separated by evidence of intense inversion tectonic movements: an older unit of sediments from Cenomanian to Early Coniacian times and a younger one from Santonian to Maastrichtian times (NIEBUHR & PROKOPH, 1997). During the inversion, the uplift of sediments of former basins or trenches was accompanied by an accelerated subsidence of the basin edges, followed by the formation of new sedimentary troughs. The inversion tectonic activity began in the centre of the Lower Saxony block with the Early Coniacian subhercynian tecto-event and influenced marginal areas during the Late Santonian - Campanian time interval.

Numerous marine transgressions and regressions occurred during latest Cretaceous times (ERNST *et al*, 1996). There is evidence of six transgressions from Santonian to Maastrichtian times, those of the Early Campanian having extended widely.

III.6.- British Islands

During Campanian times, the sea spread over the Shetland platform. In the southern part of the Viking graben, development of claystones or marls sedimentation occurred during the Coniacian - Early Campanian time span. At the northern end of the graben, nearly the whole succession is clastic.

In the Cardigan Bay - British Channel, most of the Campanian sequence comprises soft chalk with hardgrounds and flint horizons. In the Irish Sea, pelagic limestones with accessory flint bands occurred from Santonian to Maastrichtian times. In the Faeroe - Shetland basin, the relative tectonic calm of the Late Cretaceous period was disturbed by a Campanian tectonic pulse. This caused rejuvenation of existing structural highs that shed turbidites of carbonate sand into the basin. Pelagic sedimentation prevailed during

Turonian times and continued until at least Middle Campanian times in the Central North Sea.

III.7.- Western Europe

Diverse palaeoenvironmental evolutions characterised Early Campanian times in south-western Europe. Marine transgressions arose in some places while continental depositional environments settled in other places at the same time. Such facts depended likely on the north-westward moving of the Iberian plate, which probably started during latest Santonian and earliest Campanian times.

The way of this Iberian drift led both to compressive stress eastwards and to distensive ones westwards along the space between European and Iberian plates. First of all, bringing together the two plates resulted in the closing of the Cantabrian - Pyrenean trough, which previously allowed a marine connection between the Tethyan and Atlantic oceanic domains.

Compressive movements were particularly active in Provence and the Eastern Pyrenees. Continental deposits (fluviatile and lacustrine facies including lignitic beds) accumulated in Provence (Valdonnian and Fuvelian *pro parte* local stages) and in Languedoc (BABINOT *et al.*, 1983). Great and variable thicknesses suggest that the continental basins were proto-foreland basins. In the Eastern Pyrenees, coastal deposits, especially deltaic formations such as the Alet sandstones (BILOTTE, 1985), prograded from the east to the west in the subpyrenean marine trough so that they progressively filled it.

It is assumed that distensive movements kept going in the Basco-Cantabrian basin, widely open towards the Gulf of Gascogne. During Early Campanian times, the area of deposition of flysch facies extended broadly in contrast with the Santonian flysch system (MATHEY, 1986; RAZIN, 1989). Moreover, the new flysch deposits were sandy instead of calcareous. At the same time, marine transgressions covered the shelves all around the basin.

To the north, chalky limestones rich in pelagic fauna, and chert-bearing limestones, deposited on the Aquitaine outer platform (PLATEL, 1989). The Aquitaine carbonate inner platform was restricted to a narrow belt which received terrigenous supplies from coastal environments along the western rim of the Massif Central.

To the south, sandy marls and argillaceous limestones rich in ammonites and planktonic foraminifera accumulated in the subsident Cantabrian outer shelf (GRÄFE, 1994, 1998). Landward, the marine transgression allowed a carbonate platform to develop widely, as well as in the southern Pyrenees. Calcareous deposits are often very rich in rudists (such as in the Quintanaloma and Santo Domingo de Silos Formations (FLOQUET, 1991). It seems that the marine transgression was the cause of a new opening of the Iberian Strait, but only coastal and very shallow calcareous and terrigenous facies deposited in the centre of this strait. On its Tethyan side, a new carbonate platform appeared also, including rudist facies in the Prebetic domain (MARTIN CHIVELET, 1993).

Another shallow seaway, through the Armorican massif to the north-west and the Massif Central to the south-east probably allowed Atlantic and Paris marine basins to connect together.

Chalks deposited predominantly in the Paris basin, enlarged during Early Campanian times. It was broadly open north-westward to the London chalky shelf, northward to the North Sea basin and south-eastward to the Helvetic shelf and farther to the Alpine Tethyan oceanic domain.

III.8.- Maghreb

In Morocco, the Senonian palaeogeography remained practically similar to that of the Turonian, the facies changes resulting from a general regression. In the Atlasic domain, the breakdown of the Turonian shelf produced small basins where confined facies (bituminous marls) deposited. These basins were surrounded by uplifted and emerged areas.

Clastic coastal sedimentation prevailed in the Essaouira, Agadir and Tarfaya basins. Eastward, the sedimentation was dominated by lagoonal and continental facies, with frequent red sands deposit.

During Early Campanian times, the Saharan platform experienced an extensive transgression which covered a large area previously emerged or subject to a brackish sedimentation.

In Tademaït and Tinrhert areas, the Campanian series is not easily separated from the Santonian one (AMARD *et al.*, 1981). These two series are made of an alternation of marls and oolitic limestones with ostracods, miliolids, pelecypods and echinoderms, witnessing of a shallow coastal marine environment.

In Tunisia, during Campanian - Maastrichtian times, direction of extension rotated from ENE-WSW to roughly E-W. Isopach maps of Abiod Formation in Sahel and Gabès Gulf indicate a similar NW-SE to WNW-ESE orientation of the basins. North-westward, in Sahel and in the Gabès Gulf, the main Campanian - Maastrichtian structural directions are NW-SE.

In Central Tunisia, a marly and clayey sedimentation with oysters, echinids and planktonic foraminifera (Aleg Formation) prevailed. This sedimentation extended southward up to the Chott area (ABDALLAH *et al.*, 1995).

III.9.- Sirt basin

During Turonian times, the Sub-Hercynian orogenic event resulted in erosion in the central and southern parts of the Sirt basin. Because of their continued and expanding onlap, Campanian rocks cover a wider area than underlying strata (WENNEKERS *et al.*, 1996). They include mainly limestones, sandstones, minor dolomites and anhydrite beds in the southern part of the Sirt basin and more argillaceous carbonates and clays in its northern part.

Campanian strata are absent from stable structurally high areas but display continued onlap and encroachment on existing structures. These include islands in the Campanian sea and lands bordering this sea to the south and north-east. Campanian sediments did

not deposit over Al Jabal al Akhdar due to a palaeotopography associated with Santonian uplift.

III.10.- Egypt - Sinai Peninsula

The Campanian - Maastrichtian lithofacies distribution indicates an extensive transgression with a gradually rising sealevel superimposed by synsedimentary tectonics.

Late Senonian deposits comprise chalks, marls and cherts, which characterise open-sea conditions. Sedimentation was mainly controlled by a basin - swell morphology

Following a regressive period during Late Santonian times, with major emergence and erosion, a transgressive period due to an increasing subsidence occurred on the whole region during Campanian times. It created a southward facies retrogradation, with clastic sediments and subordinate carbonates. To the north, most Campanian strata are composed of massive chalks and marls, mainly formed in deeper shelf environments.

III.11.- Syria - Levant

Overall deepening and transgression of the sea occurred quite in the whole Syrian and Levant areas during Campanian times. However, this general trend took place whereas important tectonic movements affected the Syrian Arc, resulting in the formation of swells and basins. Thus, nature and thicknesses of the deposits changed according to the highs and lows location.

In Palmyrides, Early Campanian deposits are mainly marly limestones intercalated by thin limestones beds and chert lenses and nodules (middle part of the R'mah Formation within the Soukneh Group; AL MALEH & MOUTY, 1994). These deposits attest to different open sea conditions, which settled during a subsiding phase of the sea floor. In a general trend, marine environments were deeper to the north, west and south-west (abundant ammonites and planktonic foraminifera), while they were shallower to the south-east and east.

North-eastwards, faulting in the Euphrates region originated a broad N 120 E orientated graben with dramatic thickening of latest Cretaceous sediments. The main pulse of rifting occurred during Campanian times (synrift deposits are up to 1500 metres thick) (JAMAL, 1998). Early Campanian facies are mainly cherty limestones rich in planktonic foraminifera (upper part of the R'mah Formation, JAMAL, 1998; lowermost part? of the Shiranish Formation according to DE RUITER *et al.*, 1994). Farther to the north-east, the Sinjar region underwent rifting during latest Cretaceous times, while the Euphrates graben also opened, but along a roughly E-W-orientated axis.

Northward, a carbonate shelf developed on the Mardin High in southern Turkey. The southern shelf edge was almost parallel to the today Turkish - Syrian boundary. Carbonates pertaining to the Massive Limestone Formation, according to KENT & HICKMAN (1997), prograded southward from the Mardin High into the northern Syrian basins during a syn-rift phase. In the Jebel Abd Al Aziz rift (*sensu* KENT & HICKMAN, 1997) fine grained basinal mudstones accumulated. Some calcareous breccias including rudists are interbedded, presumably coming also from the Mardin High. In the Kurd Dagh and Aafrîne areas, during Early Campanian times, argillaceous mudstones plentiful of planktonic foraminifera and holding organic matter (source rocks) deposited in open sea, probably under euxinic conditions (AL MALEH, 1976). To the north, a reduced and glauconitic - phosphatic series (AL MALEH, 1976) indicates more oxygenated and shallower depositional environments, maybe pertaining to the slope that linked the outer pelagic domain to the carbonate Mardin shelf.

In Lebanon, Early Campanian deposits form the middle part of the "Chekka Formation". Made of white chalks and marly chalks rich in planktonic foraminifera, these deposits include some phosphate and chert beds accumulated in an outer shelf. This latter was directly linked westwards to the Tethyan basin. Similar sedimentary features are recorded from Israel Campanian series.

Early Campanian deposits in Jordan comprise massive bedded chert and brecciated chert interbedded with planktonic bearing limestones (part of the lower half of the Amman Formation; ABED & SADAQAH, 1998). The source of silica was biogenic at a time of high productivity related to an intense upwelling from the Tethyan basin to the west up to the open shelf to the east.

III.12.- Arabian plate

According to the recent work of HUGHES & FILATOFF (1997), a Late Cretaceous pre-rift sequence can be described west of the Arabian domain. A poorly sorted, locally kaolinitic, pebbly sandstone with minor interbedded siltstone lies unconformably on the Precambrian basement. According to its palynological content, a Campanian - Early Maastrichtian age is attributed to this unit. Palaeoenvironment corresponded to a non-marine, braided river complex, grading in the upper part, into a continental succession including marginal marine facies. Early Campanian deposits could be included in the lower part of the Mukalla Formation, which is well represented in the Yemen area and consists of current bedded, medium to coarse-grained sandstones.

Following the eoalpine compressive tectonic episode, which led to the overthrusting of the Sumeini and Hawasina nappes and to the obduction of the Samail nappe onto the margin of the Oman platform in the Early Campanian, the Arabian Peninsula once again became submerged in Late Santonian - Early Campanian times (LE MÉTOUR *et al.*, 1995). In Oman, the Suneinah foreland basin was a narrow, elongate trough, parallel to the Eoalpine Oman Mountain chain fault. It extended southward into the intra-shelf basin of Interior Oman. The trough resulted from downflexing of the Arabian platform in front of the mountain belt that was generated by the Eoalpine orogeny (LE MÉTOUR *et al.*, 1995). The trough and the intrashelf basin were the sites of accumulation of basinal hemipelagic mudstones, chalk and shales represented by the Fiqa Formation. On the high axes of the Haushi - Huqf uplift and southern Dhofar, the first marine deposits above the Precambrian basement are of Early

Campanian age (PLATEL *et al.*, 1994). They correspond to mixed rudist rich carbonates and siliciclastic facies (Samhan Formation, Fig. 15.2).

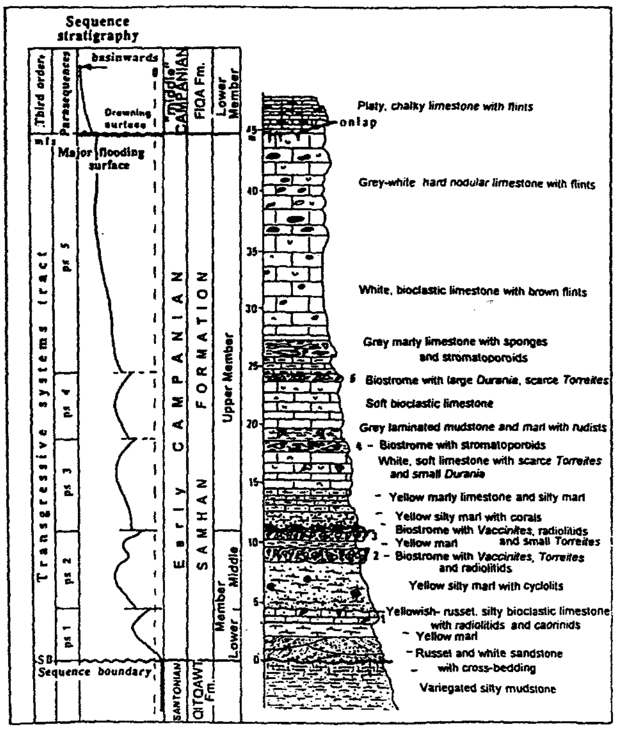

Fig. 15.2. - Lithostratigraphic column of the Early Campanian Samhan Formation in Saiwan area (eastern Oman), from PLATEL *et al.* (1994).

In Saudi Arabia, data (VASLET *et al.*, 1990) indicate that the first marine deposits overlying the continental deposits of the Upper Wasia Formation are of Campanian age (Khanasir Member of the Aruma Formation). These deposits are limestones bearing ammonites, inoceramids, echinids and dasycladals, deposited in shallow marine conditions. However, the Early Campanian age of the base of the Khanasir Member has not been, until now, formally proved.

IV.- PALAEOENVIRONMENTS AND FACIES

IV.1.- Exposed lands

On the northern margin, the Fenno-Scandian Shield and its extension to the east formed a huge cratonic domain. It was the source of clastic material, mainly deposited on the Russian platform.

To the north, the Paikhoy - Novozehlesky Range displayed a W-E elongated massif, which separated the boreal domain from the Pechorian sea and the Polar - Uralian Strait.

To the east, south of the Middle Uralian Strait, the south Uralian Range acted as a palaeogeographic barrier separating the western Siberian domain from the Russian platform.

To the south, the Ukrainian Shield and its Dniepr - Donetz satellites massifs formed highs between the Ukrainian basin and the Crimea - Scythian platform.

The Silesian massif and its south-eastern extension corresponded to the major Tornquist - Teisseyre tectonic lineament. It bounded to the east the western European domain. This latter comprised four main groups of lands: the Bohemian - Rhenish massif, the British Islands massifs, the Armorican massif and the Massif Central, the Iberian and Ebro - Corsica - Sardinia massifs.

On the southern margin, the emerged lands pertained to the wide Saharan and Arabian Shields. However, it is possible to distinguish different units.

In the Maghreb area, lands extended broadly by comparison to those of Late Cenomanian times. They corresponded roughly to W-E elongated ridges, such as the Idrissides and Jebilet Highs, bounding intracontinental basins.

A central Saharan craton was more likely a promontory between the Tademaït and Nile continental or littoral basins.

The incipient continental Red Sea graben separated the Sudan craton and Arabian Shield.

To the far-east, in Oman, Late Santonian compressive movements gave rise to a tectonic bulge, some parts of it are believed to have been emerged.

IV.2.- Fluviatile, lacustrine, fluvio-lacustrine

On the northern margin, fluvio-lacustrine environments are well documented only in southern France and north-eastern Spain.

In Provence, continental deposits that constituted the Valdonnian and Fuvelian local stages, are made of palustrine and lacustrine limestones bearing *Corbicula*, *Unio*, and marls and siltstones. Lignitic beds and seams intercalate. According to biostratigraphic data and correlations, the Valdonnian and Fuvelian have been attributed to the whole Campanian standard stage. However, magnetostratigraphic data (WESTPHAL & DURAND, 1990) correlate the Valdonnian to the end of the long C34N magnetic Chron i.e. to the Late Santonian substage and the Fuvelian to the C33R magnetic Chron i.e. to the Early Campanian substage.

In Languedoc (Saint Chinian and Boutenac areas), fluviatiles sandstones bearing *Unios* are attributed to the Campanian stage (FREYTET, 1970).

In the Maestrazgo and Valencia region, the first Late Cretaceous continental deposits, essentially made of lacustrine and palustrine limestones bearing *Lychnus ellipticus*, are probably of Early Campanian age (ALONSO *et al.*, 1993).

On the southern margin, continental deposits are known in Morocco, Libya and Egypt, and on the eastern and southern border of the Arabian Shield, especially in Yemen and Egypt.

In Morocco, continental marls including gypsiferous clays, and red sandstones accumulated in the W-E elongated pre-African trough (CHOUBERT & FAURE-MURET,

1962) located between the Anti-Atlas and Haut-Atlas highs.

Mainly red sands belonging to the Mut Formation, upper part of the Hawashiya Formation and Barmil Formation (MATEER *et al.*, 1992) deposited in the Nile basin.

A continental siliciclastic facies, regarded as a pre-rift sedimentary constitutive sequence, settled in a NNW-SSE elongated depression, approximately at the emplacement of the today Red Sea.

In central and north-western Somalia, eastern and northern Ethiopia, continental sandstones pertaining to the Jesoma Formation *pro parte* and to the Tisje Formation (MATEER *et al.*, 1992) characterise braided stream environments. In western Yemen, a terrestrial sedimentation, made of sandy facies for the most part, is recorded in the base of the Hallah Formation and in the middle part of the Mukalla Formation (MATEER *et al.*, 1992.

IV.3.- Coastal marine, shallow marine (terrigenous - clastic)

Coastal silicilastic belts rimmed almost all the emerged shields or massifs, although terrigenous supplies were particularly important on the southern border of the Russian craton connected with the Russian platform, and around the Arabian Shield.

Glauconite, radiolaria and scarce planktonic foraminifera indicate that the shallow littoral environments to the east and south of the Uralian Range were widely opened to the eastern Siberian Sea and to the Russian platform. A large deltaic system, bordered by outer diatomaceous deposits separated the Russian platform from the Ukrainian basin.

In Morocco, marls and oysters-rich limestones attest of shallow marine gulfs. Similarly, the Tademaït area corresponded to a shallow embayment (HERKAT, 1999) with terrigenous inputs and deposition of sands and oysters and echinids-rich clays and marls.

In Somalia and Yemen (Hadramaout), crossbedded sandstones and subordinate limestones and shales (BOSELLINI, 1989) indicate deltaic and shallow marine environments.

IV.4.- Carbonate platforms

Carbonate platforms are poorly represented during Early Campanian times by contrast with those figured on the Late Cenomanian map. This fact is probably related to the overall deepening and transgression occurring then.

Rudist carbonate ramps are restricted on the northern Tethyan margin to reduced areas in Aquitaine, Pyrenees and northern Spain.

On the southern margin, carbonate platforms bearing large benthonic foraminifera and rudist fragments are reported in the Cyrenaic area around Jebel Alakdar.

The presence of resedimented rudists and other benthonic components, as well as carbonate breccias, within basinal marls in northern Syria, attest to the existence of a carbonate swell to the north, related to the Mardin High.

Early Campanian rudist carbonate platform and banks, surrounded by clastic environments, settled in south-eastern Oman around presumably exposed highs such as the Huqf massif (PLATEL *et al.*, 1994).

IV.5.- Hemipelagic environments

By contrast with the restriction of the carbonate platforms, the deeper carbonates and hemipelagic oozes environments extended broadly on both Tethyan margins during Early Campanian times.

A prevalent chalky facies accumulated within these environments characterised by an accumulation of skeletons of planktonic marine algae (calcispheres, thoracospheres and *Nannoconus*). Other subordinated organisms are observed too, including planktonic and benthonic foraminifera, fragments of inoceramids, bryozoa, echinoderms (HANCOCK, 1990). One could define chalk as coccolithic limestones.

Chalk, the dominant facies in Western Europe as a whole during Late Cretaceous times, accumulated at depths of 100 to around 600 metres. In parts of the trenches of the North sea domain, such as the Viking graben and the Central graben, greater depths prevailed, possibly reaching 1000 metres. Much of the chalks probably deposited there by mass flows from the flanks of the trenches (HANCOCK, 1990). Chalks formed the reservoirs of the Dan and Ekofisk fields in this North sea domain (HANCOCK, 1990).

Landwards, the coccolith-rich carbonate mud grade to greensands, which deposited in belts around the emerged massifs. These facies are made of glauconitic-rich sands or clayey sands, possibly phosphatic and showing sedimentary gaps. The Early Campanian Vaals Greensand in Northern Belgium and South Limburg are representative of such environments (BLESS & ROBASZYNSKI, 1988).

The occurrence of siliceous facies (opoka) corresponding to abundant sponge association in the Early Campanian series, as in the Polish Trough, is indicative of shallower shelf environment.

In Crimea, Scythian and Russian platforms, deposition of chalky limestones and marls with planktonic (*Globotruncana arca*) and benthonic (*Bolivinoides decoratus*) foraminifera attests to lower sublittoral - upper bathyal environments (BARABOSHKIN, *in litteris*, 1998).

On the main part of the Saharan platform (HERKAT, 1999), planktonic foraminifera bearing limestones and clayey limestones characterise open shelf environments. Northwards, argillaceous sedimentation prevails and indicates deeper environments, especially in the Tunisian outer basins and South Atlasic troughs. Similar deeper clayed deposits prevail in Syrt grabens.

In the entire Levant areas, deep-outer shelves layered into phosphatic swells and in basins where chalky limestones accumulated.

Deep hemipelagic carbonates extended on the whole outer shelf of the Arabian Shield from the north to the south until the Fiqa basin in Emirates and Oman. Similar environments linked this basin to the south-eastern margin of the Arabian plate and to the Owen basin.

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16.- LATE MAASTRICHTIAN (69.5 - 65 Ma)

JEAN PHILIP[1] & MARC FLOQUET[1]

I.- MAIN FEATURES

Based on various time scales, the duration of the Late Maastrichtian substage is estimated at between 2.5 and 4.5 Ma (from 69 up to 66.5 Ma according to HAQ *et al.*,1987; from 69.5 up to 65 Ma according to GRADSTEIN *et al.*, 1995; from 69.42 up to 65 Ma according to HARDENBOL *et al.*, 1998). The time scale of GRADSTEIN *et al.* (1995) has been chosen for the mapping, i.e. a 4.5 Ma time span.

Based on magnetostratigraphic evidence, Late Maastrichtian times extended from the upper part of the C31 polarity chronozone up to the middle part of the C29 polarity chronozone. The K/T boundary (KTb) is located within the C29 reverse polarity chronozone (GRADSTEIN *et al.*, 1995). In the South Atlantic, the KTb has been dated at 350 ka above the base of 29R (HERBERT & D'HONDT, 1990).

In deep or open marine formations, according to HARDENBOL *et al.* (1998), Late Maastrichtian datations are generally obtained according to the distribution of ammonites (Gollevillensis Zone in Northern Europe, and Fresvillensis and Terminus Zones in Southern Europe), planktonic foraminifera (Mayaroensis Zone), belemnites (Belemnitella junior and Belemnella casimirovensis Zones) and nannoplankton (Micula murus and Micula prinsii Zones).

In shallow marine formations, accurate datations are supplied by large benthonic foraminifera (*Orbitoides media, Orbitoides apiculata, Siderolites calcitrapoides, Omphalocyclus macroporus, Lepidorbitoides socialis*), rudists (FAD of *Hippurites cornucopiae*, FAD of *Dictyoptychus morgani*, LAD of *Hippuritella laperousei*, etc.) and ostracods.

In continental formations, charophytes (*Peckichara* sp.1, *Microchara cristata*...), gasteropoda (*Lychnus giganteus, Lychnus bourguignati*...) etc., provide reliable biostratigraphic data.

Late Maastrichtian sea-level changes corresponded successively to: 1) a slight sea level fall, 2) a conspicuous sea level rise and 3) a marked sea level drop followed by a moderate sea level rise until the end of the Cretaceous (HARDENBOL *et al.*, 1998).

Middle - Late Maastrichtian times are thought to correspond to a transitional period between an equatorial - dominated thermohaline oceanic circulation and a polar - dominated one, inducing a sharp decrease in organic productivity (MOUNT *et al.*, 1986). Such a reversal, around the Middle - Late Maastrichtian boundary (71 – 70 Ma), is considered as controlled by sea level changes (BARRERA *et al.*, 1997).

At the Middle - Late Maastrichtian boundary (70 Ma), stable isotope records of ocean temperature and salinity attest to cooler and less saline bottom water in oceans of the southern hemisphere, and the coeval occurrence of widespread biotic changes (MACLEOD & HUBER, 1996). In this respect, FRANCK & ARTHUR (1999) have recently pointed out to the tectonic forcings of Maastrichtian climate evolution. The onset of Laramide tectonism during Middle Maastrichtian times led later to the concurrent draining of major epicontinental seaways. Together with the major reorganisation of the oceanic circulation, these events resulted in cooling and increasing latitudinal temperature gradients and ventilation of deep oceans, thus affecting a range of marine biota, such as rudists and inoceramids.

II.- STRUCTURAL SETTING AND KINEMATICS

The positions of the Laurasian and African megaplates are deduced from the location of the palaeomagnetic pole given by BESSE & COURTILLOT (1991).

In the former alpine oceanic, Apulia and Eurasia kept coming closer, giving rise to local collisions and subsequent highs such as the Austroalpine and Dacide block. After the obduction of ophiolites during Late Santonian - Early Campanian times along the margin of Arabia, the remaining oceanic domain continued to disappear during Late Maastrichtian times, beneath the overthrusting of the Pontides, Kirsehir, and Sanandadj - Sirjan block. At the same moment the enlargement of two oceanic domains occurred at: 1) the South Caspian - Eastern and Western Black Sea and 2) the Eastern Mediterranean basin.

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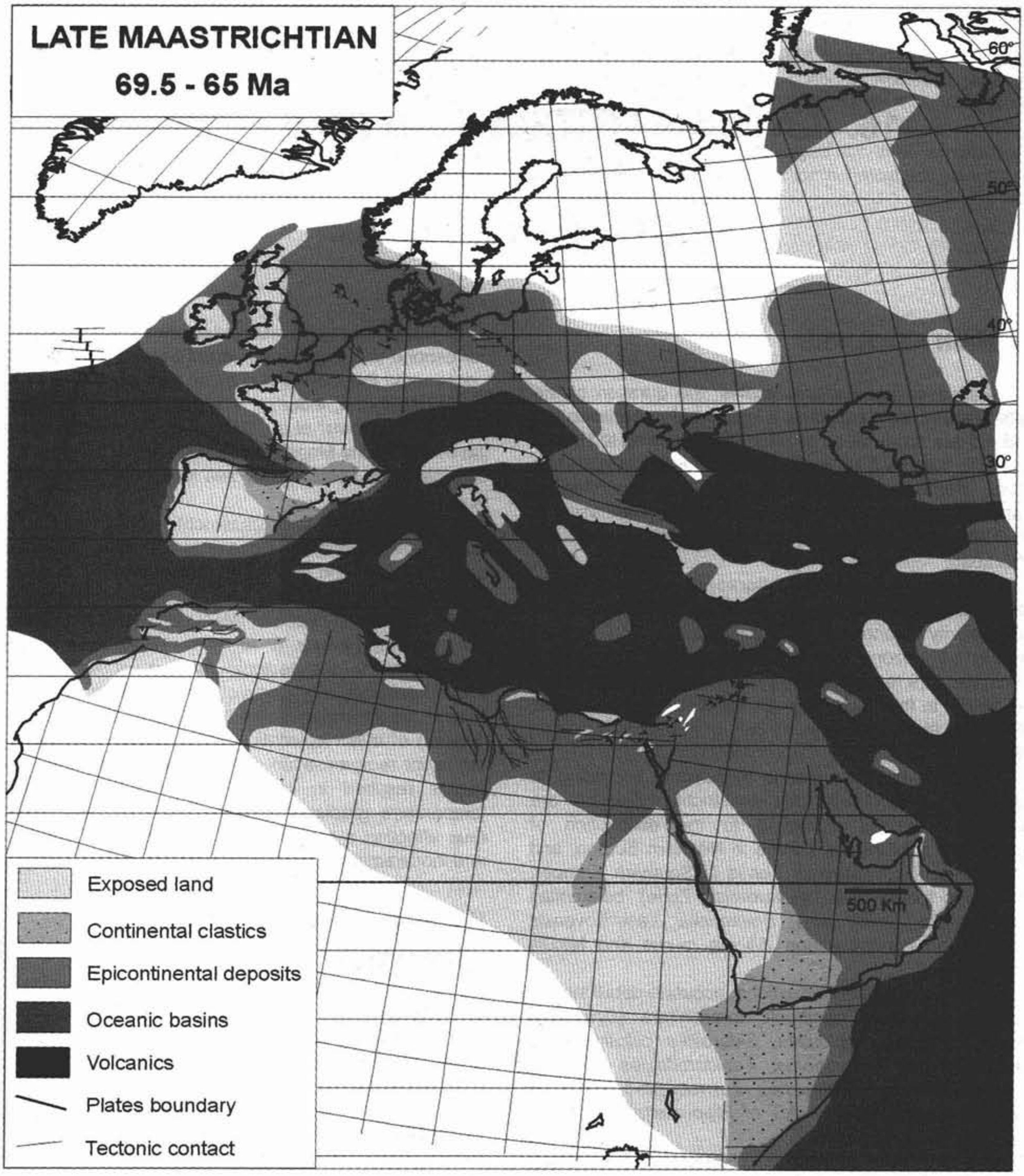

Fig. 16.1: Simplified palaeogeographic map of Peri-Tethyan area during the Late Maastrichtian.

III.- Definition of domains

III.1.- Turan, Mangyshlak, Ural

Maastrichtian deposits in the north-western part of the Turan plate comprise brown sandy ferruginous clays.

Late Maastrichtian interval in the Turgay area is characterised by *Belemnella casimirovensis* (Amon *et al.*, 1997). From Coniacian up to Maastrichtian times, a white chalk unit sedimented over the Mangyshlak area (Gaetani *et al.*, 1998). Late Maastrichtian deposits show firstly regressive characters, then a brief transgressive trend, so called "elegans transgression" (Kopaevich *et al.*, 1999) and finally a marked regression which coincided with the sea level fall generally recognised close to the Maastrichtian/Danian boundary.

A Middle Uralian Strait joined the Russian platform and the west Siberian Sea. This strait separated the South Uralian exposed land and the North - Middle Uralian Range.

Like during Early Campanian times, the polar Uralian Strait linked the west Siberian Sea to the Pechorian shelf. The Paikhoy - Novozemelsky Range probably remained exposed during Late Maastrichtian times and limited this shelf to the north.

III.2.- Russian platform, Ukraine, Scythian - Crimean platform

The identification of Late Maastrichtian deposits in this area is supported mainly by benthonic foraminifera *Brotzenella preacuta* and *Hanzawaia ekblomi*, belemnitellid *Belemnella casimirovensis* and by calcareous nannofossils *Nephrolithes frequens*. Correlation with the Maastrichtian series of north-western Europe is thus possible.

Late Maastrichtian times corresponded to a period of overall regression on the Russian platform. Thus, the western coastline of the platform moved southward on several hundred kilometres (ALEKSEEV, *in litteris*, 1999). However, open marine depositional environments of outer shelf developed widely in the Precaspian depression, in the Volga river region, in the Dniepr - Donetz depression, and to the south in the Scythian platform. Such environments are recorded in prevailing carbonate marly sedimentation. The Dniepr - Donetz marine areas were connected to the west to the Polish trough through the Ukrainian basin. In this area, mainly marls and sandy-marls, represent Late Maastrichtian deposits, except in its western part where siltstones and opokas intercalated.

The Ukrainian Shield and Donetz High united in one large block of low land as a result of tectonic uplift (ALEKSEEV, *in litteris*, 1999). Southwards from this block a shallow and warm marine basin existed in South-western Crimea. At the opposite, a deep-water bathyal basin remained in Eastern Crimea.

III.3.- Moesian platform and Black Sea

To the north of the Balkan thrust belt, an outer shelf marine carbonate sedimentation occurred, possibly with deeper environment to the north-west (Carpathian Flysch trough) and to the east (Black Sea) and with irregular clastic content in Western Moesia and Southern Carpathian areas. To the south of the Balkan Thrust Belt, the rift basin was almost filled in and terrigenous - clastic deposits dominated (GEORGIEV, *in litteris*, 2000).

As previously, the western Black Sea experienced during Maastrichtian times a post-rift regime resulting in deposits of mixed volcanics and chalky sediments.

In the eastern Black Sea, petroleum drillings have shown an apparently complete Late Cretaceous to Danian pre-rift sequence of turbidites and chalks unconformably overlain by Late Eocene mudstones. North of the today Turkish coast, Late Cretaceous (Late Maastrichtian?) volcanics, tuffs and chalks including tuffaceous materials, are overlain unconformably by Eocene mudstones.

III.4.- The Polish Trough

During Late Maastrichtian - Early Tertiary times, the tectonic inversion of the Polish trough occurred and the Mid Polish Swell originated in a transpressional regime (LAMARCHE *et al.*, 1998).

Oscillations of transgressive and regressive impulses were characteristic features of the Polish Trough during Maastrichtian times. Thus, the general trend of the basinal domain was to a gradual deepening and extension north-eastwards during late Early Maastrichtian times. Its maximum extent was reached on the turn to early Late Maastrichtian times. Both during Early Maastrichtian and early Late Maastrichtian times there was a clear domination of limestones, chalk and opokas sedimentation. Later during Late Maastrichtian times, the sea was gradually shoaling, and opokas, marls, gaizes, and sandstones deposits increased accordingly. In the southern part of the Polish trough, marly chalks and opokas are attributed to the *Belemnitella junior* and *Hoploscaphites constrictus crassus* Late Maastrichtian Zones (BLASKIEWICZ, 1980).

III.5.- Northern European platform

Chalk sedimentation with flint interbeddings prevailed on the Northern European platform. Late Maastrichtian times are well documented in Lower Saxony and on the Pompeckj Block (Hemmoor area; SCHMID, 1982) and in the classical Stevns Klint type locality in Denmark. Late Maastrichtian deposits are at least 85 metres thick in the area of Lägerdorf - Kronsmoor - Hemmoor where the Early - Late Maastrichtian is marked by the occurrence of *Belemnitella* gr. *junior* (SCHULZ *et al.*, 1983).

In the intra-shelf basins of northern Germany (Lower Saxony, Münster basin), ERNST & KÜCHLER (1992) recognised a Middle Maastrichtian eustatic - event (sea-level fall around -71 Ma) and a so called "junior transgression" during Late Maastrichtian times, which is considered by these authors as triggered both by eustasy and tectonic incidents such as fault block movements.

III.6.- British Islands and North Sea

Within the Cardigan Bay - British Channel area, interbedded soft chalk and hardgrounds / flint horizons developed during latest Campanian and Maastrichtian times. The original depositional limit of the chalk is not well defined; lack of significant clastic material suggests that land was distant, or that elevations were low. The Highland Border Ridge to the north of the Irish Sea was submerged only during Late Campanian times and probably up to Maastrichtian times. It is likely that thin Late Cretaceous (Late Maastrichtian?) limestone and chalk were deposited extensively over much of Ireland and the Irish Sea area and, may be, on the present onshore eastern border of the British Islands. However, the transitional deposits between the North Sea chalky basins and the exposed lands during Maastrichtian times are poorly known due to Tertiary erosional processes.

A change of the tectonic regime occurred in some places on the turn of the Cretaceous - Tertiary boundary. Some areas that had accumulated a thick sequence of sediments now reversed the process and were uplifted (HANCOCK, 1990). Such uplifts are known as inversion axes. In the Central graben, the uplift was completed before Late Palaeocene times and likely began as early as Late Maastrichtian times. During Late Campanian to Maastrichtian times, limestones and chalks were deposited in the Central graben. Outside this unit, Maas-

trichtian sedimentation consisted in bioturbated pelagic facies.

III.7.- Western Europe

The SE-NW orientated French craton was formed by linkage of the Massif Central and Armorican massif and hindered communications between the north-eastern European chalky domain and the south-western Atlantic shallow siliciclastic and carbonated domain. This craton developed widely and was rimmed by broad coastal belts mainly constituted of clastic deposits. Consequently, the Paris chalky strait contracted between this craton and the Rhenish - Bohemian massif. Actually, preserved Late Maastrichtian chalky facies are very scarce within the Paris basin. They are located especially in its north-western (QUESNEL *et al.*, 1996) and northern parts, this latter corresponding to the northern extension of the strait, which was opened towards the North Sea basin.

In the Maastricht type area, chalky calcarenitic sedimentation developed up to the Late Maastrichtian Meersen Member. A conspicuous hard ground, the Berg en Teblijt Horizon, has been interpreted as fitting with the K/T boundary in this area (SMIT & BRINKHUIS, 1996).

An overall marine regression characterised Late Maastrichtian times in south-western Europe. The general regression was mainly due to tectonic compressive movements generated by the north-westward displacement of the Iberian plate (including its north-eastern and northern promontories i.e. the Corso-Sarde massif and the Briançonnais; STAMPFLI, 1993). Owing to this Iberian plate drifting, compressional events resulted in diachronous emersions.

The previous Cantabrian - Pyrenean - Provençal trough closed step by step from the east to the west: as early as in the course of latest Santonian times in Provence, and during Campanian up to Maastrichtian times along the Pyrenees. The previous Iberian strait also progressively disappeared during Campanian and Maastrichtian times, due to coeval marine retreats south-eastward and north-westward from a central hinge located in the Soria Province (FLOQUET, 1991).

Instead, continental basins developed, acting as proto-foreland basins. Fluvio-lacustrine and palustrine basins are reported in Provence (as recorded, for example, by the Rognac Limestone) and in Languedoc, in Eastern and Central Pyrenees, in northern Iberian Ranges (as recorded in part of the Santibañez del Val Formation; FLOQUET *et al.*, 1994) and in the southern Iberian Ranges as well (BABINOT *et al.*, 1983). Consequently, as soon as during Late Maastrichtian times, the French Craton, the Corso-Sarde and Ebro massif, and the Iberian massif joined together in order to form a single and broad land.

Seas were restricted in the Alpine Tethyan domain to the east, in the Prebetic and Betic Tethyan domains to the south, and in the western Pyrenean and Biscay Atlantic domains to the west.

Widespread deep-water flysch facies deposited along the western and northern Tethyan margin in the western Alps. On the contrary, flysch deposits were minor in the western Pyrenean trough between the Aquitaine and Basco-Cantabrian margins. In this basin, *Zoophycos* -bearing shaly and marly deposits replaced progressively during Maastrichtian times the sandy flysch facies (MATHEY, 1986 ; RAZIN, 1989).

Landwards, carbonate platforms developed only in the form of narrow belts, including scarce rudist banks, such as the Aquitaine platform, the Cantabrian and Prebetic platforms. The sedimentary general trend was that of marine carbonates followed by sandy and argillaceous deposits belonging to nearshore, lagoonal and marshy environments (Tremp Formation in the Southern Pyrenees, GALLEMI *et al.*, 1983 ; Auzas Marls Formation in the northern Pyrenees; BILOTTE, 1985). However, intercalations of Orbitoids rich sandy limestones (such as the Nankin Limestones) recorded minor sea-level changes and slight marine transgressions.

III.8.- Maghreb, Malta Escarpment

In Morocco, Maastrichtian series generally rest unconformably on prior formations and are overlain unconformably by Eocene deposits (CHOUBERT & FAURE-MURET, 1962). The sedimentation was shallow marine with predominance of terrigenous facies (marls) and occurrence of first phosphatic deposits.

Mixed shallow marine and brackish environments, with alternation of dasycladals rich limestones and gypsiferous clays developed in the southern part of the Saharan platform (Tinrhert). However, in this area and in Tademaït, Late Maastrichtian times are characterised by marine limestones bearing echinids, pelecypods, nautiles (AMARD *et al.*, 1981). These limestones progressively evolved to similar facies bearing numerous *Laffiteina*, probably of Early Palaeocene age. During Maastrichtian times, the Saharan platform was progressively covered by a carbonate sedimentation (Abiod Formation of Tunisia) characterised by chalky limestones bearing large foraminifera (Orbitoids), echinids and cephalopods. According to BUROLLET (1956), these deposits bear witness to a shallow and open marine environment. At the end of the Maastrichtian stage, a well-marked clayey organic rich sedimentation occurred (Argiles d'El Haria), in Southern Tunisia, in the Gafsa area (BUROLLET, 1956), and in the Tunisian trough. The Cretaceous-Tertiary boundary lies in these clayey deposits. A recent study of the Cretaceous - Tertiary transition in the Gafsa area (KELLER *et al.*, 1998) has shown that during the last 300 Ka of the Maastrichtian stage, deposition occurred in a inner neritic (littoral) environment that shallowed to a nearshore hyposaline and hypoxic environment during the last 100 - 200 Ka of Maastrichtian times.

Limestones and argillaceous limestones of the El Haria Formation were penetrated by wells in the western part of the Malta Escarpment (BISHOP & DEBONO, 1996).

III.9.- Sirt basin

The end of Late Cretaceous times saw the maximum extent of the marine transgression (WENNEKERS *et al.*, 1996). Maastrichtian series are widespread, being thin or absent at a few scattered localities but, only on the horst blocks. A mixture of clastics and carbonates make up the section and they potentially form excellent reservoirs especially around palaeotopographic highs where highly

energetic proximal skeletal carbonates and calcarenites are concentrated.

Generally, deep marine conditions resulted in deposition of the widespread transgressive Sirte Shale. This formation and the Kalash one are equivalent facies. The distal Kalash Member deposited in structurally low areas while the proximal Kalash Member (limestones) accumulated fringing islands in the Maastrichtian sea. The basal contact is unconformable with the upper section of the Campanian Rachmat Formation. Contemporaneous fault movements can be documented thanks to thick deposits of over 500 m in trough areas. The main attribute of this succession is that it has proven to be the major source for the bulk of the hydrocarbons in the Sirt basin. Extremely rich kerogen content is present in the Sirte Shale attesting to deposition in an oxygen depleted marine reducing environment (WENNEKERS *et al.*, 1996).

III.10.- Egypt - Sinai Peninsula

Late Cretaceous times corresponded to the onset of the closure of the Neo-Tethyan Ocean. As a consequence, a structural inversion occurred along the former extensional regimes of the unstable shelves. Right lateral transpression and compressional tectonics during Senonian times caused inversion along the deep-seated faults that developed during Mesozoic rifting in the region. The area is affected by small scale E-W (Eastern Desert and Northern Sinai) and NE-SW (Southern Israel) trending swell/basin morphologies during Maastrichtian times. The extension events were dominated by two main perpendicular syn-depositional extensional trends, NNE-SSW and WNW-ESE respectively (MOSTAFA, 1999). The period between Maastrichtian and Early Eocene stages recorded almost no sedimentation, as a result of the uplifting during the folding process.

In Egypt, the Maastrichtian transgression reached further to the south (compared to those of earlier Cretaceous periods) and covered most areas with thick chalk-marl units, shales, and subordinate silts of the deeper shelf (KUSS & BACHMANN, 1996). Lateral and vertical changes of lithofacies patterns and thickness are the result of a complex interplay of regional subsidence, pronounced eustatic sea level changes and differential sedimentation rates across the region during Maastrichtian times.

III.11.- Syria - Levant

A major marine transgression, initiated during Late Campanian times, reached its maximum during Middle and Late Maastrichtian times allowing the sea to flood the main part of the Arabian peninsula. Syria and Levant (Israel, Lebanon, Palestine and main parts of Jordan), at the north-western edge of this peninsula, pertained to a deep outer shelf, where chalky limestones and argillaceous and marly limestones deposited. Facies are very rich in planktonic foraminifera (especially globotruncanids and heterohelicids) and Calcisphaerulidae (as represented by the Syrian Shiranish Formation *sensu lato*, Mount Scopus Group in Israel, Ghareb Formation in Jordan).

However, isolated sedimentary basins formed, separated by highs, likely governed by early plicative deformations, which gave rise to the Syrian Arc (SHAHAR, 1994).

The Syrian and Levantine outer shelf domains were opened both to the north and to the west, towards the eastern Mediterranean oceanic basin. Carbonate platforms developed farther to the south-east, on the middle and inner shelf.

In Syria, the Euphrates graben was acting until Early - Middle Maastrichtian times. During the Late Maastrichtian substage the rifting ended definitively and the previous normal faulting began to invert owing to N-S and NNW-SSE compressive stress (JAMAL, 1998). Compressive movements also occurred in the Palmyrides.

Such movements were due to the overall convergent trend linked to the closing of the northern Neo-Tethyan ocean and to the obduction which took place along the northern border of the Arabian plate from Turkey to Oman. In addition, close to the Maastrichtian - Palaeocene boundary, sharp sedimentary discontinuities are recorded in pelagic series and may express a marked sea level fall (eustatic?) occurrence at that time.

III.12.- Arabian plate

West of the Arabian domain, in the Saudi Arabia Red Sea, siltstones and interbedded mature sandstones representing a coastal and deltaic/estuarine environment can be referred to latest Maastrichtian times (HUGHES & FILATOFF, 1997).

The last transgressive episode of the Cretaceous period that flooded extensively the Arabian platform occurred during Late Maastrichtian times.

In southern Yemen areas, the Maastrichtian Sharwain Formation is formed by marls, shales and calcareous marls grading to the west to a more detrital, clastic regime (ALSHARHAN & NAIRN, 1997).

In Oman, Maastrichtian transgressive deposits unconformably overlie either the Campanian succession or the older substratum (LE MÉTOUR *et al.*, 1995). In Dhofar and Haushi - Huqf areas Late Maastrichtian times are recorded by mixed carbonates and siliciclastics bearing large foraminifera (*Omphalocyclus*, *Loftusia*) and rudists. In the Oman mountains, the Maastrichtian succession most commonly comprises the basal clastic unit of the Qahlah Formation (of Late Campanian to Early Maastrichtian age) which is overlain by the upper carbonate unit of the Simsima Formation. Shallow carbonate facies grade transitionally to the open marine facies of the Sunainah trough. In some regions (Jebel Ja'alan, Batain Coast), the Simsima Formation is sharply overlain by Late Maastrichtian debris flows deposits (ROGER *et al.*, 1998).

At the scale of the long period of time (4 Ma), from Late Maastrichtian up to Danian times, the Oman Mountains recorded profound modifications linked to probable uplift of the Arabian platform and to the flexuration of its margins (ROGER *et al.*, 1998.). In many areas, the Cretaceous/Tertiary boundary corresponds to a major break in sedimentation, and local emergence.

In Saudi Arabia, rudist rich carbonate facies developed during Late Maastrichtian times. These facies constitute the Hajajah Member and the lower part of

the Lina Member of the Aruma Formation (El Asa'ad, 1983).

III.13.- Alpine Tethyan domain

Palaeoenvironments and palaeogeographic units of this domain have been reported according to the previous Late Maastrichtian map from Camoin *et al.* (1993). In addition, some precisions have been drawn from Poisson (*in litteris*, 1999) for the Anatolian microcontinent and the surrounding areas.

An accretionary prism was formed at the emplacement of the previous Neo-Tethys ocean and above its northward subducted part. The former volcanic arc remained active and became emergent from place to place. Detritic platforms developed around these highs, separated from the Anatolian microcontinent (Taurus platforms) by a pelagic basin. A part of the northern Neo-Tethys oceanic crust was obducted onto the margins of these platforms while the main part of them remained the site of shallow marine carbonate sedimentation.

In Munzur Dagh, shallow marine carbonates of Late Maastrichtian age cap ophiolites. These latter took their origin from a northerly-located Neo-Tethyan basin (northern Neo-Tethys). The ophiolites were obducted while this basin closed probably during Early Maastrichtian times.

Very shallow marine carbonates (Liburnian facies) developed on the Beysehir and Bey Daglari isolated platforms. Planktonic carbonates accumulated around these platforms, especially in the Menderes area.

To the south of the Anatolian microcontinent, the closure of the Pamphylian basin finished and was the site of plankton ooze carbonate sedimentation and, to the east, of flysch deposition. Farther to the west, the eastern Mediterranean Herodotus deep basin reached its largest extent.

IV.- Palaeoenvironments and facies

IV.1.- Exposed lands

On the northern margin, a huge emerged land played a prominent role since it began to extend from the Fenno-Scandian Shield up to the Uralian Range. This Range formed the western boundary of the West Siberian domain. However, restricted communications remained between this domain and the Russian platform through the Middle Uralian Strait.

Isolated exposed lands or massifs existed in the northern Tethyan Shelf and separated distinct areas of marine sedimentation. The Donetz High plus the Ukrainian Shield acted as a barrier between, on one hand, the Ukrainian basin and the Polish Trough and, on the other hand, the Scythian - Crimean - Moesian platforms.

The Silesian massif and its south-eastern extension were probably implied in the tectonic inversion that began during Late Maastrichtian times.

The Bohemian and Rhenish massifs played the same palaeogeographic role while separating the northern European basins and the northern Alpine shelves.

Ascertained exposed lands were located to the south-west of the European shelves and belonged to the Massif Central, Ebro massif, Corsica - Sardinian block and Iberian massif. Tectonic Laramide uplifts probably originated these continental blocks.

On the southern margin, the Saharan and Arabian exposed Shields were entered by local straits, which were situated in vicinity of Tademaït, Nile and Red Sea present areas.

Late Cretaceous tectonics gave rise to local highs in Morocco (Idrissides High), Tunisia (Kasserine High), Libya (Jebel Alakhdar High) and Egypt.

The Proto-Oman Mountains, generated from the Campanian ophiolites obduction, acted as a palaeogeographic barrier between the Arabian platform and the eastern Neo-Tethyan ocean.

IV.2.- Fluviatile, lacustrine, fluvio-lacustrine

Lacustrine and palustrine limestones, alternating with fluviatile clays and sands, deposited in Late Maastrichtian continental basins of the Western European margin. These basins formed in a compressive tectonic setting corresponding to the Laramian phase.

In Provence and Languedoc areas, palustrine and lacustrine environments are recorded in the Rognac limestones bearing gastropoda and pelecypods. Evidence of periodic emergences are given by numerous palaeosoils. Pebbly facies and red clayed sands, in some places very rich in dinosaur remains and oncolites, are characteristical of braided rivers channels and floodplains. Locally, foothills facies, such as breccias, witness to synsedimentary tectonics that gave rise to the Laramian foldings and subsequent intracontinental basins.

Analogous facies, environments and tectonics are known in diverse continental basins, which developed at that time between the Iberian and Ebro massifs (Fig. 16.2).

Such depositional environments also occurred on the southern margin, in Morocco, Sudan (Nile basin), south-eastern part of the Arabian Shield, and in Ethiopia - Somalia.

In North-Western Sudan, the Kababish Formation reflects continental to transitional marine environments (Mateer *et al.*, 1992).

In Northern Yemen, the upper part of the Ghiras Formation, made of sandy to pebbly deposits with thin siltstones and mudstones, is typical of braided rivers and overbanks environments (Alsharhan & Nairn, 1997).

In Northern Somalia and Eastern Ethiopia, the Yesomma sandstones represent fluviatile environments (Bosellini, 1989). South of Berbera, this formation rests unconformably upon Late Jurassic limestones and is arranged in fining upward sequences topped by massive mudstones and shales. Crossbedding directions indicate an eastward and south-eastward transport (Bosellini, 1989).

IV.3.- Coastal marine, shallow marine

Terrigenous sandy-clayed deposits prevailed on the margin of the Fenno-Scandian Shield and Uralian Ranges. These deposits probably existed on the European continental shelves and are to be related to the Northern and Western European exposed massifs.

Sands and sandy marls including glauconite of open marine coastal environments deposited along the eastern and northern borders of the north - middle Ural Range. To the west and east of the Russian platform, existed a coastal belt of clastic sediments (sands and sandy marls) of shallow marine environments (ALEKSEEV, *in litteris*, 1999).

In Aquitaine, northern Pyrenees and northern Spain, coastal environments are best represented by quartzose bioclastic limestones bearing red algae, orbitoids, echinids, etc. Locally in the southern Cantabrian basin, these facies graded to oysters bearing sandstones and calcareous sandstones indicative of shoreline (Fig. 16.2) (FLOQUET, 1991; PLUCHERY, 1995). In South-Eastern Spain (Altiplano de Jumilla - Yecla), analogous facies, which constitute the Calizas Arenosas del Molar Formation (MARTIN CHIVELET, 1993) are typical of nearshore environments.

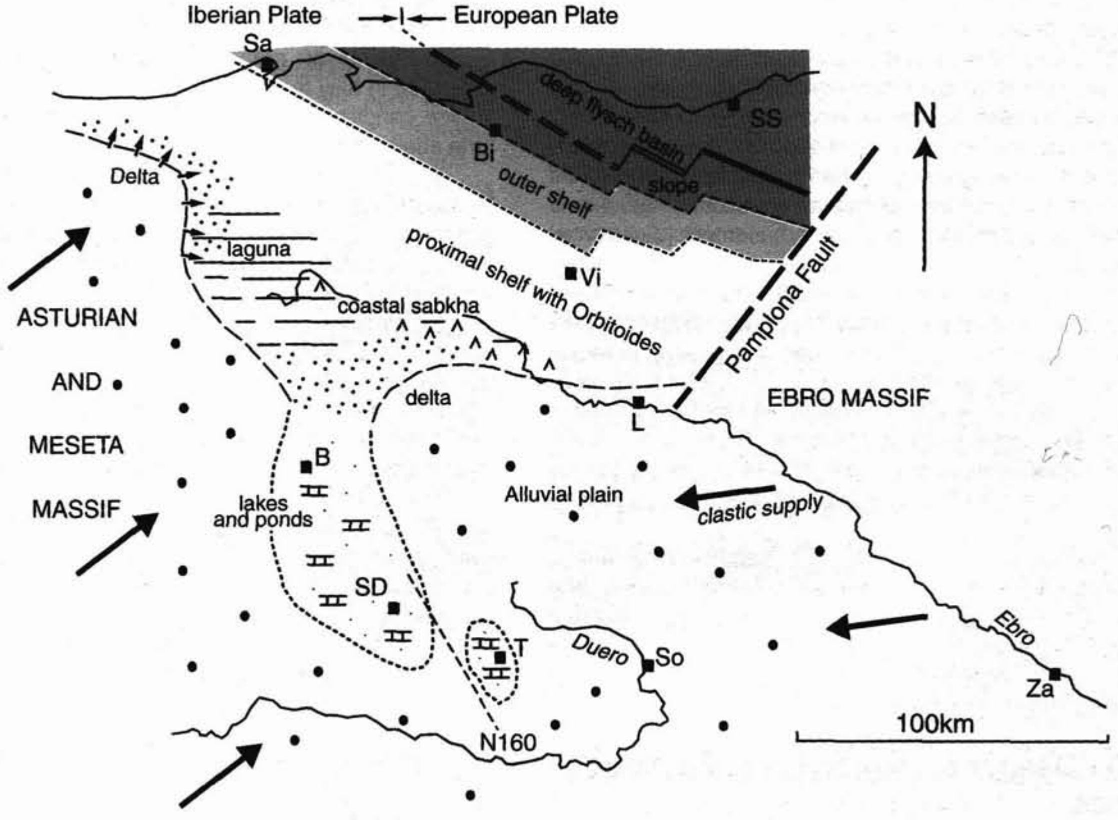

Fig. 16.2: Palaeogeography of the northern Castilian coastal complex including continental fluvio-lacustrine and alluvial basins, during Late Maastrichtian times, from FLOQUET *et al.* (1994). B, Burgos; Bi, Bilbao; Ln, Logroño; Sa, Santander; SS, San Sebastián; So, Soria; Vi, Vitoria; Za, Zaragoza; SD, Santo Domingo de Silos basin; T, Arganza - Talveila - Cubilla basin.

On the southern margin, terrigenous coastal marine deposits accumulated mainly in Morocco, Tademaït, Southern Egypt, around the Arabian Shield and in Somalia.

In Morocco, clastic facies plus marls and siliceous limestones associated to phosphatic deposits, in some places rich in fishes and reptiles debris, characterise shallow open marine environments.

In Tademaït - Tinrhert area, mixed coastal marine and shallow marine environments with fluctuating salinities existed.

In Southern Egypt and northern Sudan, on the northern margin of the Nubian Shield, coastal environments with siliciclastic sedimentation probably settled owing to the Late Maastrichtian transgression (FLEXER & REYMENT, 1989).

In Somalia, dark clays with scattered lignitic layers occurred, interbedded with sandstones, in a N-S elongated belt representing coastal environments. These deposits evolved eastwards into calcareous shales of shallow open marine environments (BOSELLINI, 1989).

IV.4.- Carbonate platforms

Carbonate platforms bearing rudist banks are poorly recorded on the northern margin. For example, only scattered Radiolitids and Hippuritids associated with large

benthonic foraminifera are known in the latest Maastrichtian Meersen Member in the Limburg area.

On the southern margin, wide carbonate platforms developed in the Saharan and Libyan areas and the Arabian Shield, mainly as a response to the Late Maastrichtian transgression.

In South-Eastern Algeria and Southern Tunisia, bioclastic limestones rich in large foraminifera (*Orbitoides, Omphalocyclus, Siderolites* and *Laffiteina*) and oysters, bryozoa, echinids, represent a shallow marine carbonate environment.

In Western Libya and Sirt area, limestones bearing large foraminifera such as *Omphalocyclus, Siderolites* and *Orbitoides,* and rudists, oysters and echinoids characterised analogous shallow water carbonate environments (BARR, 1972; ASHOUR, 1996).

Environments of carbonate platform-bearing rudists and same large foraminifera are typified by the Aruma formation in Saudi Arabia.

Around the Proto-Oman mountains, Late Maastrichtian rudist banks grading upward to bioclastic limestones rich in large foraminifera (Simsima Formation) onlaped onto the obducted ophiolitic - radiolaritic basement (SKELTON *et al.*, 1990).

In Somalia, a belt of bioclastic wackestones-packstones and shally grainstones - boundstones, referred to a Maastrichtian - Palaeocene succession, is regarded as deposits of shallow carbonate platform environments by BOSELLINI (1989). Locally, in Northern Somalia, rudists and corals bearing limestones (PONS *et al.*, 1992) could be attributed to a Late Maastrichtian carbonate platform owing to the presence of the genus *Dictyoptychus*.

In the Apulian domain and neighbouring areas, extensional tectonic movements created a horst and graben mosaic which governed the setting of isolated carbonate platforms (CAMOIN *et al.*, 1993). Sedimentologic and environmental features of these carbonate platforms have been summarised by these authors.

IV.5.- Deeper carbonates, hemipelagic oozes

Hemipelagic environment deposits, including limestones or chalks, sometimes mixed with clays, fine sands, cherts, are well recorded on the northern margin where they extend on large areas. The regions involved in this environment are from east to west: the Turan, Russian and Scythian platforms, Moesia, the Polish Trough, Low Saxony, the Danish trough, the North Sea basin and neighbouring domains, the Celtic Sea, and the Gulf of Gascogne margin. Recent investigations (QUESNEL *et al.*, 1996) have confirmed the deposition of chalk with flints until the Late Maastrichtian in the western Paris basin.

Nektonic (belemnites) and pelagic (planktonic foraminifera, nannoplankton) organisms are well represented in these facies and bear witness to an open marine to relatively deep environment. In addition, benthonic faunas (bivalves, brachiopods, bryozoans, echinids) and frequent burrows, indicate normal conditions of sea bottom oxygenation.

On the southern margin, hemipelagic deposits are represented in Western Morocco, and in a wide and relatively continuous belt of outer shelf deposits, which runs from North-Western Algeria to Eastern Oman.

Hemipelagic deposits entered the Morocco margin by the gulf of Agadir. Chalks and marls were deposited at low subsidence and sedimentation rates. The high contents of siliceous and calcareous plankton and of phosphatic concretions suggest high plankton fertility.

In the Libyan areas, mainly calcareous mudstones deposited in fully marine outer neritic environments. Deep marine conditions are inferred from the deposition of the Sirte shales and the Kalash argillaceous calcilutites.

In Southern Tunisia, clays inputs during Late Maastrichtian times were linked to a seasonal warm to temperate climate that changed to warm/humid conditions with high rainfall. Decreasing surface productivity is also documented (KELLER *et al.*, 1998).

Hemipelagic deep sea chalks and clays, locally mixed with siliciclastic limestones, chert bands or nodules, and marls, broadly deposited on the huge outer shelf located to the north of the Arabian Shield. Local turbiditic inputs reflect uplift and erosion on elevated highs formed by synsedimentary tectonics (KUSS & BACHMANN, 1996).

In the Oman foredeep basin (Aruma basin), sedimentation of basinal turbiditic clastics prevailed. These clastics (sands, shales, conglomerates) were provided by the obducted Proto-Oman Mountains. Far from the exposed lands (i.e. the South-Western Iran outer shelf), thinly bedded limestones and argillaceous limestones containing planktonic foraminifera (Gurpi Formation) deposited.

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MAPS 17 TO 23 - TERTIARY

J.E. MEULENKAMP[1] & W. SISSINGH[1]

INTRODUCTION

Seven maps portray the Tertiary palaeogeographical and environmental settings of the Peri-Tethys domains for the Early Eocene (Early to Middle Ypresian), Middle Eocene (Late Lutetian), late Early Oligocene (Late Rupelian), late Early Miocene (Early Burdigalian), early Middle Miocene (Early Langhian), Late Miocene (Late Tortonian) and Middle/Late Pliocene (Piacenzian/Gelasian). Three of the maps have counterparts for the Tethyan domains (DERCOURT *et al.*, 1993); no such maps were published for the Early Ypresian, the Early Burdigalian, the Early Langhian and the Middle/Late Pliocene. Consequently, no Tethyan data are included in the Ypresian map, while only major palaeogeographical/palaeoenvironmental and/or tectonic features are presented for the Tethyan domains in the Early Langhian and Piacenzian / Gelasian maps. The Early Burdigalian configuration was partly taken from the Late Burdigalian Tethys map.

For each time-slice, simplified palaeogeographic maps are included in the text. On these maps "epicontinental" deposits refer to sequences laid down in shallow-marine environments or in shallow-water environments with salinities deviating from normal.

The palaeogeographical - palaeoenvironmental maps reflect the effects of large-scale inversion which affected the Peri-Tethyan platforms during the Tertiary in response to geodynamic processes in the Tethyan realm. These processes were related to large-scale temporal and spatial shifts in oceanic subduction and subduction roll-back along the collision zone, as well as to the (partial) incorporation of Tethyan domains in the adjacent cratons. The continent - continent collision became increasingly effective through enhanced mechanical coupling of the African/Apulian and Eurasian continental lithospheric plates since about the Middle Eocene - Late Eocene transition, around 40 Ma ago. This ultimately resulted in a strong fragmentation of, in particular, the northern Peri-Tethys platform, which, in turn, caused an increasing differentiation between palaeogeographical and biogeographical domains in the course of time.

One of the major episodes of differentiation straddles the Eocene - Oligocene transition, about 34 Ma ago. At that time, the northern and southern domains of the hitherto existing comprehensive Tethyan realm became palaeogeographically and biogeographically separated through the incipient emergence of the Alpine chains. These northern domains and the adjacent parts of the northern platform are commonly referred to as the Paratethys.

CHRONOSTRATIGRAPHY AND GEOCHRONOLOGY

The time-progressive isolation of basins and the concomitant overall shift from open marine to brackish and fluvio-lacustrine sedimentation conditions on the Peri-Tethys platforms and in the Tethys - Peri-Tethys transitional zones in the course of the Tertiary resulted in the development of endemic faunas and floras in various parts of the palaeogeographically reconstructed domains. Because of this, first-order correlations of many of the Tertiary sedimentary sequences with the standard planktonic foraminiferal and calcareous nannoplankton zones are often impeded or impossible. As a consequence, regional biostratigraphic scales and regional stages have been established for, for instance, the Central/Western and the Eastern Paratethys and the Paris basin. In addition, an independent chronostratigraphic/biochronological framework, based on evolutionary trends and migration events of micro- and macromammal associations, exists for the Cainozoic terrestrial record. The correlation between the various bio-and chronostratigraphic scales and their calibration with numerical ages is presented in figure 17.1.

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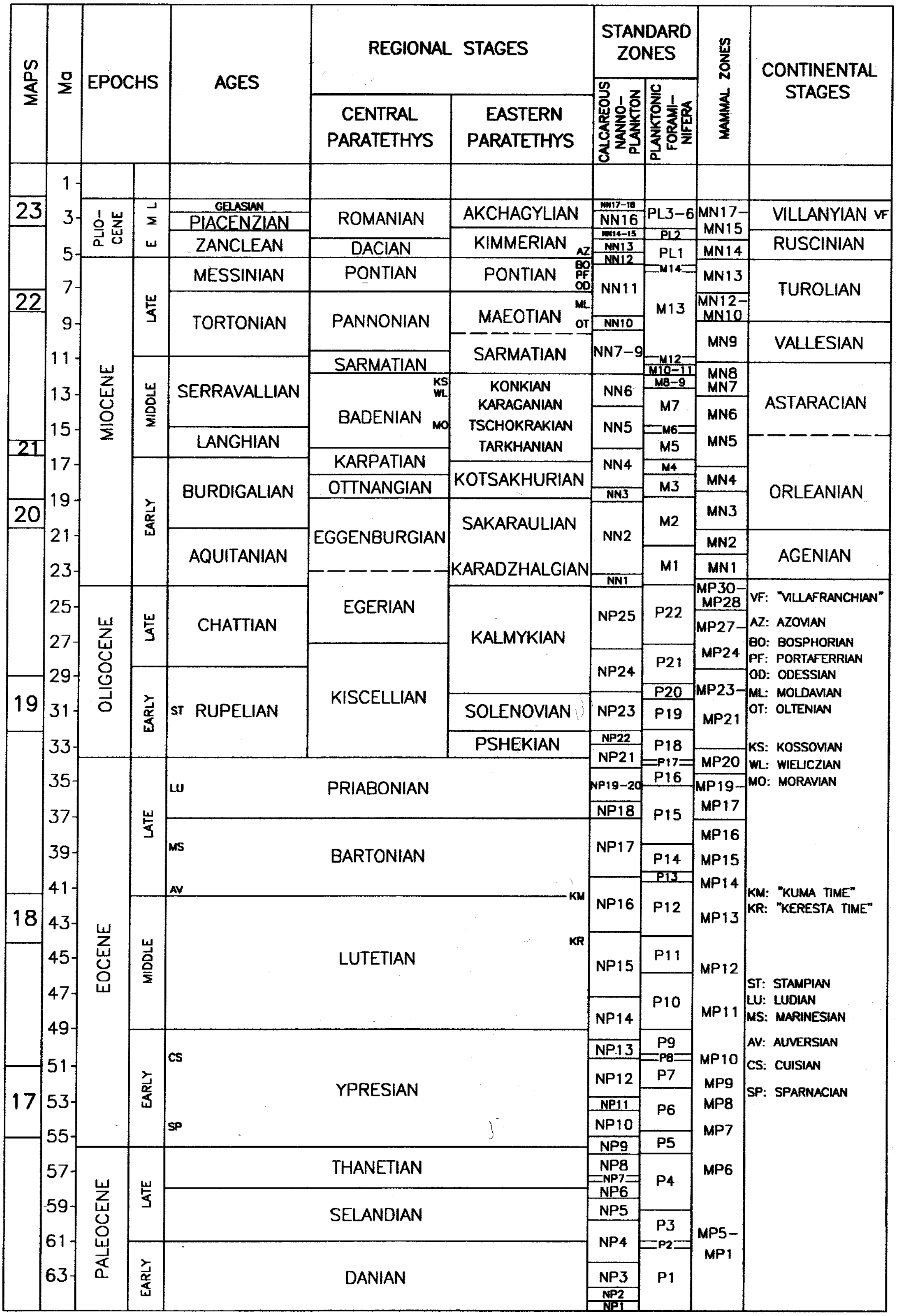

Fig. 17.1: Correlation chart of Tertiary epochs and ages with Paratethys regional (sub)stages, continental stages and marine and continental biozones. Composed after BERGGREN *et al.* (1995), RÖGL (1998, 1999) and MARUNTEANU (1992, 1999). Position of Palaeogene mammal zones after LEGENDRE & LÉVÊQUE (1997); Paris basin (sub)stages after GÉLY & LORENZ (1991).

17.- EARLY TO MIDDLE YPRESIAN (55 - 51 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The Early to Middle Ypresian is represented in the sedimentary record by the (upper part of) the planktonic foraminiferal standard zones P6 and P7, which are correlative to the calcareous nannoplankton zones N10 to N12 (Fig. 17.1). In the Crimean - Caucasian domains the interval comprises the regional *Morozovella subbotinae* and *M. marginodentata* zones. The Early to Middle Ypresian time interval thus covers the interval from ~55 Ma to ~51 Ma. In the northern part of the Iberian domain the Early Ypresian is roughly equivalent with the Ilerdian regional stage. The base of the latter stage corresponds to the base of the *Morozovella velascoensis* zone, which, in turn, is correlative to the appearance of the larger foraminiferal genera *Alveolina* and *Nummulites*. The upper limit of the Early Ypresian is roughly equivalent to the Ilerdian - Cuisian transition, marked by the boundary between the *A. trempina* and *A. oblonga* biozones. With respect to mammal biostratigraphy, the Early to Middle Ypresian is characterised by the faunal associations of zones MP7 and MP8/9 (Fig. 17.1).

I.2.- Structural setting and kinematics

In terms of the geodynamic evolution of the Tethyan - Peri-Tethyan realms, the Early Eocene may be considered intermediate between the Late Cretaceous major reorganisation of plate boundaries and the Middle to Late Eocene inception of (the major phase of) continent - continent collision, induced by the further convergence of the African/Apulian and Eurasian plates. This resulted in the uplift of large parts of the northern and southern Peri-Tethys platforms. The Iberian block with its core of Palaeozoic, Hercynian terranes (Iberian massif) acted as an "independent" lithospheric microplate for at least part of the Cainozoic. Relative motions of the African and Eurasian plates resulted in the formation of several Alpine domains and mountain ranges around the Iberian massif (Cantabrian mountains, Pyrenees, Iberian and Betic ranges) in the course of the Tertiary. Early Eocene compressional forces were transmitted to the entire Iberian block (SANZ DE GALDEANO, 1996), resulting in nappe piling and in a roughly W-E orientation of foreland basins at either side of the evolving chain of the Pyrenees, as well as in reactivation of Late Hercynian faults intersecting the Iberian massif. This, in turn, caused the separation of the Duero and Tagus basins.

Oceanic subduction was the dominant process controlling the overall northward progradation of thrusting and foredeep development with flysch accumulation in basins adjacent to the incipient island arc / mountain chains from the present Alpine domains in the west towards the present-day Carpathian domains in the east. However, the Early to Middle Ypresian palinspastic configurations are poorly known, which implies that the positions of the structures delimiting the northern Peri-Tethys platform to the south as indicated on the map are highly speculative. Deep basins existed in the Tethyan and transitional Tethyan - Peri-Tethyan domains further to the east (Black Sea depressions, Greater Caucasian basin), bordered to the south by the Iran - Lesser Caucasian volcanic belt. Palaeogeography and sedimentation along the northern margins of the southern Peri-Tethys platform were, at least in part, controlled by SW-NE trending structures related to the development of the Syrian arc from the Levantine marginal areas towards the west. On the northern Peri-Tethys platform proper the development of the Bray and Weald anticlines equally witnesses of the transmission of compressional stress from the collisional Tethys domain to the platform areas in the north.

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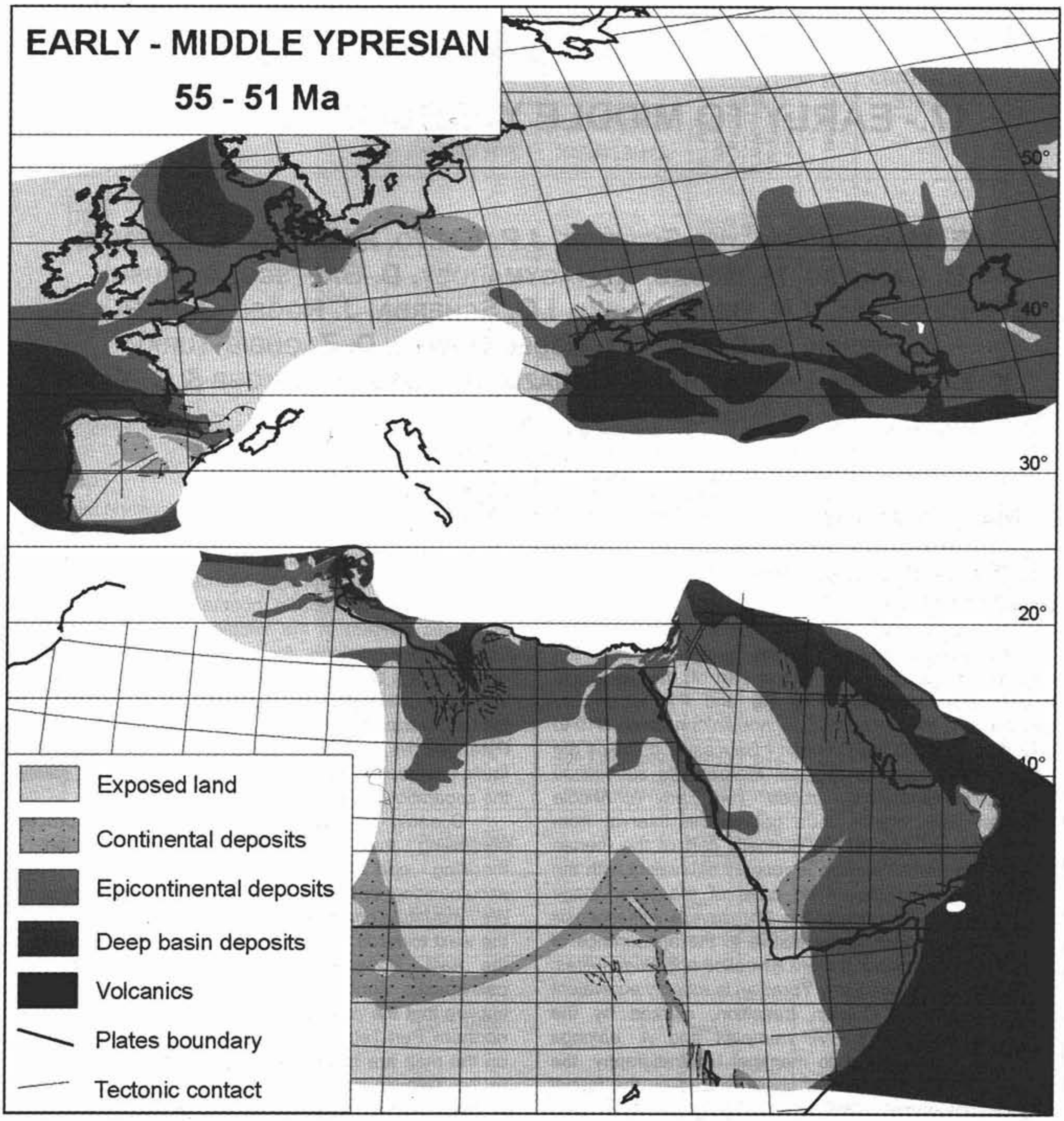

Fig. 17.2: Simplified palaeogeographic map of the Peri-Tethyan area during the Early to Middle Ypresian.

I.3.- Outlines of palaeogeography and palaeoenvironments

Large parts of both the northern and southern Peri-Tethys platforms emerged after Late Cretaceous (Maastrichtian) time in response to African - Eurasian convergence. However, various marine corridors across the northern platform persisted in the Early Palaeogene, connecting the Atlantic and Tethyan/Indo-Pacific domains. In the Early to Middle Ypresian, the marine corridor across the Aquitaine basin still connected the Atlantic Ocean with the western part of the Tethyan realm. A N-S trending seaway in the east allowed exchange of water masses between the Arctic Ocean and the Tethyan/Indo-Pacific domains via the West Siberian basin and the Turgaj Strait. The marine Atlantic - Tethys/Indo-Pacific trans-European corridor from the North Sea basin via the Polish Lowland basin and the Dniepr - Donetz depression played an important role with respect to faunal and floral distribution in the Palaeocene. This connection was temporarily disrupted in the Early Ypresian, probably in response to a eustacy-controlled regression at the end of the Palaeocene which resulted in the closure of the Pripyat Strait and, consequently, in different faunal developments in the basins at either side of the (temporarily) closed corridor (BENIAMOVSKII, 1998).

In the basins of the north-eastern parts of the northern platform a distinct faunal / floral differentiation between northern and southern domains existed in the Early to Middle Ypresian, characterised by abundant siliceous planktonic groups and organic-walled phyto-plankton, and calcareous planktonic associations (calcareous nannoplankton, foraminifera), respectively. Dysoxic to anoxic conditions were widespread in outer shelf and deeper parts of various basins of the northern Peri-Tethys realm, not only in the east, but also, for instance, in the North Sea basin. In contrast, fluvio-lacustrine and brackish sequences accumulated in the Paris basin, which was still connected with the North Sea basin at the time. In the south-west, carbonate platforms developed at either side of the sub-Pyrenean foredeep. In the basins located within the Iberian massif deposition of fluvio-lacustrine, partly evaporitic successions predominated. Carbonate platform development and accumulation of deeper-marine calcareous muds largely characterised sedimentation in the east, in the domains of the Turan Sea and the Terek - Caspian depression. On the southern Peri-Tethys platform carbonates were deposited over large areas of Arabia. Deposition of marls and shales, partly reflecting dysoxic to anoxic environments, was predominant in the shelf areas covering the north-eastern parts of the African plate.

II.- DESCRIPTION OF DOMAINS

II.1.- South-western Europe

II.1.1.- Iberian Peninsula

In the Early Eocene, marine sedimentation in the Peri-Tethys domains was confined to a W-E oriented, southward migrating trough in the Pyrenees region, in which predominantly calcareous turbidites and marls accumulated. This foreland basin was bordered by large carbonate platforms with distal ramps and erosive talus protruding into the basin. Some deltaic complexes developed along the basin margins; the bordering emerged areas provided siliciclastic deposits which formed alluvial fans. The large interior Duero, Tagus and Ebro continental basins received mainly alluvial sediments, fringing the margins of the basins. Large evaporitic depositional systems developed in the Tagus basin, whereas the fill of the Duero basin mainly consists of lacustrine carbonates. The peripheral areas of the Iberian massif in the west and in the east were subject to the deposition of mainly siliciclastic, non-marine successions.

I.1.2.- Aquitaine basin

Foredeep subsidence persisted along the northern margins of the incipient Pyrenees. Concomitantly, the Aquitaine basin was subject to transgression. However, some intra-basinal shoals existed; they correspond to anticlinal ridges (e.g., those of Villagrains, Audignon and Roquefort). Limestones, marls and sandstones accumulated in neritic to deep-water environments in the southern part of the Aquitaine basin (e.g., "Marnes de Gan", "Marnes du Louts", "Grès de Coudures" formations). In the north-west, marine, neritic marls and limestones accumulated, while the continental "Formation de Guizengeard inférieure" was deposited in the north-east (SZTRÀKOS *et al.*, 1998). In the east, a marine corridor connected the Aquitaine basin with the Tethyan domain; platform carbonates rich in tropical larger foraminiferal associations (alveolines, nummulites) were formed at either side of the deep-water corridor. Continental sedimentation prevailed in the north-eastern part of the Aquitaine basin, as evidenced by, for instance, the siderolithic sands of Charentes, the "Molasse du Libournais" and the "Molasse de l'Agenais" (DUBREUILH, 1989). The Massif Central was the provenance area of these clastics.

II.2.- Western Europe

II.2.1.- Paris basin

In latest Thanetian time, the Paris basin emerged and was subjected to large-scale erosion. However, basin subsidence and deposition was already resumed at about the beginning of the Ypresian (Sparnacian). Initially, continental and lagoonal clastic deposits (marls, clays, lignites) accumulated. Shallow-marine deposits were laid down in the northern part of the basin (especially east of the emerged Bray anticline). In the course of the Ypresian (Cuisian), the pre-existing brackish-water, lagoonal environments enlarged into a much wider, E-W trending area, which more or less crossed the central part of the basin. This environmentally restricted area of sedimentation graded northwards into a widespread shallow-marine realm that was merged with the marine North Sea basin. Accumulation of sandy continental clastics occurred in the south. These erosion products were derived from relatively low-lying palaeoreliefs including the Massif Central (MÉGNIEN, 1980; GÉLY & LORENZ, 1991).

II.2.2.- Southern North Sea basin

In general, the Tertiary sediments of the southern(most) part of the North Sea basin (Belgium, southern Netherlands) accumulated cyclically under strong influence of eustatic fluctuations in sea level and episodic uplifting of the clastic provenance area corresponding to the Ardennes - Brabant area. The sediments of the Early and Middle Ypresian are represented by a sequence consisting of sands (Basal Dongen Sand) which are overlain by clays (Ieper Clay). The sequence reflects a marine transgression which onlapped and eventually overstepped the Artois axis to the south, after a brief period of uplift and emergence of this large anticlinal structure in between the Paris basin and the North Sea basin during the latest Thanetian. Further to the north in the basin, the earliest Ypresian is represented by a tuffaceous marker bed (Basal Dongen Tuffite) which represents the acme of Thulean volcanism in the region of the Rockall - Faroe trough. The marine transgression peaked during the Middle Ypresian. Regressive depositional circumstances developed at the end of the Ypresian. They are mainly reflected by sandy deposits (oldest Brussels Sand) which were largely derived from the Rhenish massif and deposited in a relatively shallow sea open to the north (LETSCH & SISSINGH, 1983; VINKEN, 1988; VANDENBERGHE *et al.*, 1998).

II.2.3.- Alpine foreland basins *(not mapped)*

As far as known, deposition in predominantly deep-water facies occurred (almost) continuously in the Alpine foredeep, in association with regional, initially shallow-marine onlap of the Alpine foreland in northerly and westerly directions from the development of the "Palaeocene Unconformity" onwards. (ALLEN *et al.*, 1991). The hiatus corresponds to the erosional surface at the base of the Tertiary cover that was induced by overall S-N trending intra-plate compression, uplift and deformation in advance of the evolving Alpine thrust belt. It developed diachronously from Middle / Late Thanetian time until the Early Rupelian. Throughout this depositional period, a transgressive and deepening succession of continental sandstones, neritic limestones and increasingly deeper-water marls and flysch was laid down on top of the Tertiary cover basal unconformity. Concurrently with the coeval first-order onlapping and overstepping of the basin margin, cyclic transgressive - regressive deposition took place towards the north and the west from the Early Ypresian onwards (HERB, 1988). The large-scale process of time-transgressive coastal marine deposition coincided with a similarly orientated migration of the deeper-marine facies belts, including Alpine flysch with interstratified chaotic wildflysch formations derived from the evolving Alpine orogen. The basal hiatus increases correspondingly in magnitude towards the external parts of the northern and western foredeep basins. The associated tectonic uplift of the foreland led to emergence and sub-aerial exposure and to localised deposition of lacustrine limestones and continental siderolithic clastics, which iron-rich and kaolinitic composition reflects tropical weathering conditions. Overall, marine submergence seems to have increased in rate during the Eocene and to have been accompanied by syn-depositional flexural normal faulting (HERB, 1988; LIHOU, 1995; MENKVELD-GFELLER, 1995).

Deposition of the Ypresian and younger neritic sediments of the Alpine foredeep was generally controlled by transgressive phases, which were followed by regressions causing limited erosion. At the migratory distal basin margins, the overall time-transgressive drowning of the basal shore-face sandstone - carbonate complex can be related to tectonically-induced, rapid basin subsidence and an increased input of fine-grained terrigenous clastics representing the most distal facies of the Alpine flysch. These clastics accumulated at great water depths in the proximal parts of a generally underfilled, flexural foredeep basin, in front of the advancing Alpine orogenic wedge.

During the Late Ypresian, the Bauges basin developed as a shallow-marine gulf that extended northward from the Alpine foredeep proper into the region comprising the present-day Bonnes plateau, the Aravis chain and the Bauges and Platé massifs. In particular, nummulite limestones accumulated in this basin.

II.3.- Central Europe

II.3.1.- Polish Lowland basin

After the regression at the end of the Early Palaeocene, the Polish Lowland domain was subjected to erosion. Both the Late Palaeocene and Early Eocene (Ypresian) successions witness of deposition in predominantly terrestrial and fresh-water (limnic and marshy) environments. In the latter, marshy environments, coal accumulated in the west, near Szczecin (Zielonczyn Formation).

II.3.2.- Carpatho-Pannonian region *(not mapped)*

II.3.2.1.- General features

The present-day Carpathian - Pannonian region consists of the Carpathian orogenic belt and the Pannonian back-arc system (Fig. 17.3). Along the northern, eastern and (partly) southern margins of the mobile Alcapa and Tisza - Dacia terranes the development of the orogenic belt was accompanied by intense folding and thrusting from the Middle Eocene onward. This "Neo-Alpine" evolution was probably related to subduction of either oceanic or thinned continental crust of the Outer Carpathian flysch troughs below the Alcapa and Tisza terranes (microplates), which, in turn, caused the outward propagation of nappe piles of the Carpathian accretionary prism and the formation of flexural foredeep basins along the front of the evolving mountain chain (KOVAC *et al.*, 1998). During the Palaeogene, the "Eocene - Oligocene basins" situated on top of the Alcapa and (Tisza-) Dacia microplates developed either on the active margins of the moving plates (fore-arc basins, such as the Central Carpathian Palaeogene basins), or in an intra-plate position (e.g., Transylvanian basin, North Hungarian, Palaeogene "Buda" basin). The superimposed Pannonian basin system was formed on top of the Alcapa and Tisza - Dacia units during the Early to Middle Miocene, in response to back-arc extension. Large-scale strike slip displacements and low-angle normal faulting played an important role during the formation of the various Neogene subbasins. The concurrent crustal stretching was accompanied by calc-alkaline volcanism. The subsequent filling-up of the Pannonian basin in the course of the Late Neogene was accompanied by the uplift and emergence of the Carpatho-Pannonian domain.

Cretaceous to Palaeocene flysch sedimentation in front of incipient (island) chains of the embryonic Eastern Alps and Carpathians continued into Early Eocene (Ypresian) times. Two major systems of flysch basins and troughs can be distinguished. The basins of the present-day external units of the Western Carpathian flysch belt (e.g., Silesian, Subsilesian, Dukla and Skole units) were located relatively close to the Northern and Eastern European platforms; they passed into the basins of the Moldavide belt of the Eastern Carpathians (RAKUS *et al.*, 1990; SANDULESCU, 1988). Further to the south, the flysch trough system of the present-day Magura unit developed. This system passed towards the west into the sedimentation area of the Alpine Rheno-danubian flysch belt (POTFAJ, 1998). Probably, the two Carpathian flysch basin systems were separated by the uplifted "Silesian Cordillera" (e.g., KSIAZKIEWICZ *in* ANDRUSOV, 1965).

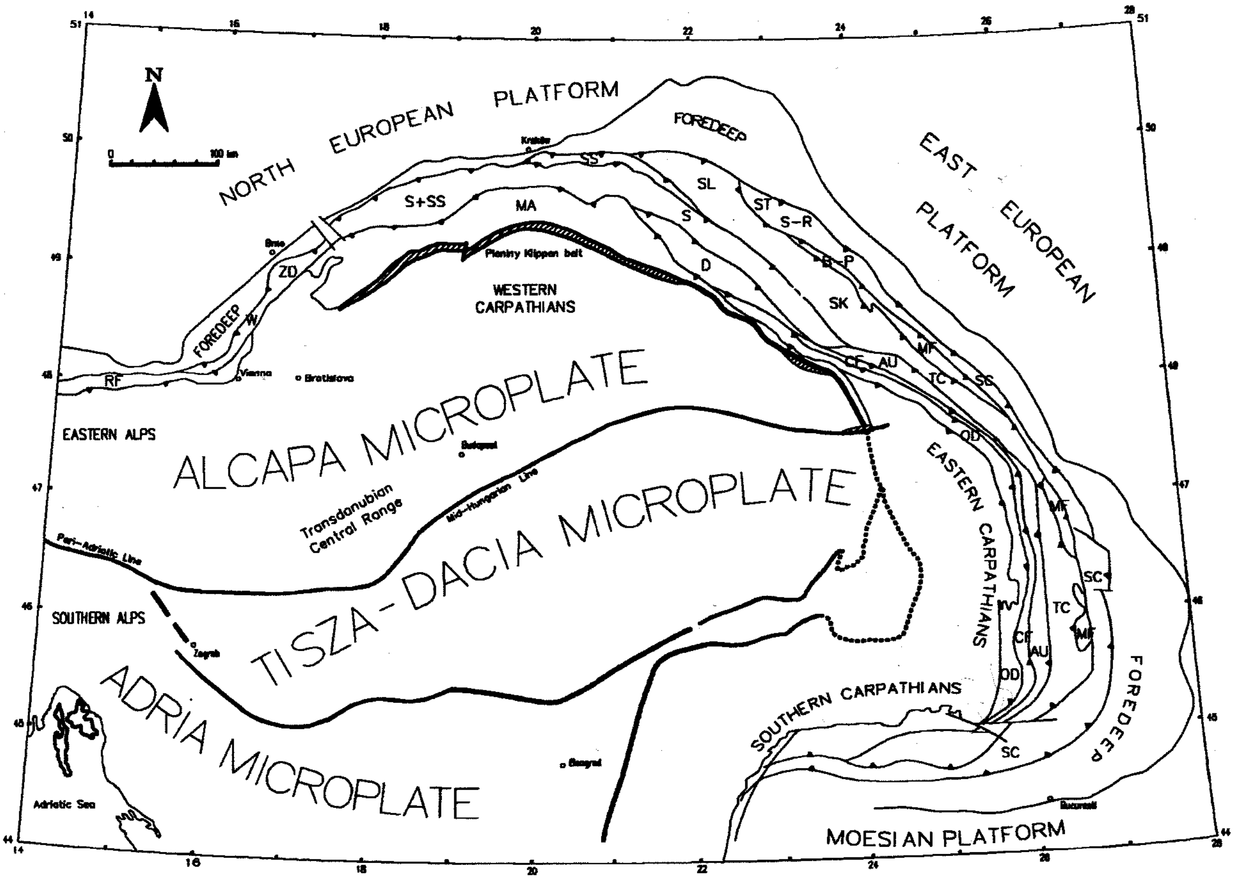

Fig. 17.3: Present-day position of the eastern Alpine and Outer Carpathian nappe units and of domains belonging to the Adria, Alpaca and Tisza - Dacia microplates (after KOVAC *et al.*, 1998).
RF, Rhenodanubian flysch; W, Waschberg zone; Z, Zdanice; SS, Subsilezian; S, Silezian; MA, Magura; D, Dukla; SL, Skole; ST, Stebnik; S-R, Sambor-Rozniatov; SK, Skiba; B-P, Borislav-Pokut; CF, Convolute flysch; AU, Audia; TC, Tarcau; OD, Outer Dacides; MF, Marginal folds; SC, Subcarpathian.

II.3.2.2.- Regional aspects

The Early Eocene sediment successions of the Western Carpathian flysch basins located close to the platform ("Krosno - Menilite zone") are characterised by variegated clays (ANDRUSOV, 1965), reflecting pelagic sedimentation. Gravity-induced sliding of slope sediments (e.g., Skole unit) derived from northern sources (RAKUS *et al.*, 1990) witnesses of ephemeral interruptions of pelagic sedimentation. Along the southern margins of the Western Carpathian flysch basins, predominantly sandstones were deposited along with various types of flysch sediments (e.g., Ciezkowice sandstone of the Silesian unit). Sediment transport directions corroborate the existence of a "Silesian Cordillera" as provenance area of clastics for the northern slopes of the Magura basin (Solan beds of the Raca subunit). Here, various types of flysch sequences were deposited, characterised by massive sandstones in the north and by thin-bedded turbidites and clays in the south (STRANIK, *in* KOLOMINSKY, 1994). In the latter part (Bystrica, Biele Karpaty and Krynica subunits), greywacke detritus was supplied from the south-east by longitudinal currents (RAKUS *et al.*, 1990). To a very limited extent, the Early Eocene sea also invaded the domain of the active margin of the Western Carpathians. Late Palaeocene to Ypresian bioherms, which originated in shallow-marine, littoral environments along the southern margin of the depositional realm of the flysch (ANDRUSOV, 1965), occur in the "Klippen belt zone". The carbonates include clasts of ("Carpathian") Mesozoic limestones. The Early Eocene successions show great differences in lithology between the inner parts of the Magura zone (greywackes) and the Central Carpathian near-Klippen belt zone (exclusively carbonates). Possibly, these differences are due to the existence of a "Magura" or "Pieniny" Cordillera separating the two sedimentation areas. However, they may as well be explained by post-depositional, Late Eocene displacement of the Central Western Carpathians relative to the flysch belt. Some small occurrences of red beds (found below Lutetian transgressive sediments) are known from the Central Western Carpathians. They are correlative to the bauxites of the Gant Formation of the Transdanubian Central Range in northern Hungary (CSASZAR, 1997).

Palinspastic reconstructions inferred from the Early Eocene sequences of the present-day south-eastern part of the Carpathians and the Transylvanian basin suggest that the Early to Middle Ypresian basin systems were situated between emerged segments of the "embryonic" Alpine mountain system ("K1 and K2 tectogenesis"; SANDULESCU, 1984) in the west and the platform domains in the east and south-east. The main, internally-located provenance areas corresponded to the Median and

External Dacides (emerged in response to the mid-Cretaceous and Laramian orogeneses). The Moesian and Scythian platforms supplied clastics from the east.

The Magura basin system did not extend into the domain of the eastern Outer Carpathians. The flysch basins of the latter domain had a width of approximately 150 km in the southern and central zones and of about 100 km in the north, as inferred from balanced cross-sections (ELLOUZ *et al.*, 1996). Sedimentation rates in these basins increased from north to south and from east to west (BADESCU, 1998). Both longitudinal and transverse facies changes occurred. The internal depositional environments were characterised by accumulation of sandy flysch with "Carpathian" debris, whereas the external domains were subject to the deposition of flysch ("Couches à hiéroglyphes") containing Dobrogean-type greenschist clasts. The deep(er)-marine, well-oxygenated flysch basin was connected with the Lom depression in the south via the Getic depression and the external shelf domains of the Moesian platform. There was no direct marine connection with the southern part of the Moldavian platform (as suggested by SAULEA *et al.*, 1971), because of the existence of an intermediary emerged area, which supplied greenschist-derived clastics. The Moldavian platform and the southern Dobrogea region formed part of the shelf zones of the Western Black Sea depression, in which glauconitic deposits with Tethyan nummulite faunas accumulated. Continental red beds (Jibou Beds) are widely distributed along the western margin of the present-day Transylvanian basin (HOSU, 1999).

There was no significant sedimentation in the Intra-Carpathian domains. On the Alcapa microplate (Pelso unit) deposition of bauxites continued; the bauxites are overlain by Early Lutetian marine deposits (CSASZAR, 1997). On the northern margin of the Tisza microplate red, grey and variegated deep-sea clays were deposited in the fore-arc basin of the Szolnok flysch trough (NAGYMAROSY & BÁLDI-BEKE, 1993; NAGYMAROSY, 1998).

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

The northern part of the north-eastern Peri-Tethys platform included the East European and West Siberian platforms and the domains of the Scythian and Turan plates. The southern parts belonged, or were transitional to, the Alpine orogenic belt (marginal basins of the Tethyan realm, Black Sea depressions, Greater Caucasian basin and Iran - Lesser Caucasian volcanic belt with associated inter-arc basins). Here, various and strongly different depositional environments existed, including areas with flysch sedimentation. In Palaeocene time, the Greater Caucasian basin was characterised by steep continental slopes and concomitant accumulation of turbidites and olistostromes (KOPP & SCHERBA, 1998). However, this pronounced differential relief smoothened during the Early Eocene. By that time the north-eastern platform parts were typified by the occurrence of shallow, epicontinental seas and emerged land masses of various extension. Pronounced differences existed between the northern and southern areas of these platform areas, as respectively marked by siliciclastic and calcareous depositional environments. This subdivision corresponded to the predominance of siliceous planktonic groups (diatoms, radiolaria) and organic-walled phytoplankton in the north, and of calcareous nannoplankton and foraminiferal associations in the south. Dark, clayey beds rich in organic matter and with aberrant faunas witness of recurrent episodes of anoxic conditions in outer shelf and deeper basinal areas during the Early to Middle Ypresian.

Wide marine corridors connecting the basin systems of the north-eastern Peri-Tethys domains with the Arctic Ocean and the tropical Tethyan realm persisted throughout the Early Eocene, as follows from the similarities in faunal and floral associations (KOZLOVA, 1993; ORESCHKINA *et al.*, 1998). The ephemeral interruption of marine sedimentation in the Pripyat depression and the north-western part of the Dniepr - Donetz basin portrayed the effects of the temporary closure of the marine corridor between the epicontinental domains of the East European platform and the North Sea basin (BENIAMOVSKII, 1998) in response to the terminal Palaeocene regression. This interruption of marine communication persisted during the Early Eocene, as evidenced by the absence of marine sediments in the Pripyat and South Baltic depressions and by the great differences in the composition of benthic foraminiferal associations at either side of the barrier (BENIAMOVSKII, 1998). Climatic proxy records (fossil leaves) indicate a change from relatively humid towards relatively arid conditions at the end of the Palaeocene. Arid floras were widespread on the East European platform and on the territories of the Turan plate from the Early Ypresian onward.

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

After its regression in latest Palaeocene time, the sea invaded again the Dniepr - Donetz depression during the Early Ypresian. Contemporaneously, marine depositional conditions occurred in most of the domains of the Scythian plate and the eastern parts of the East European platform, as far as the present-day Middle Volga. In the shallow-water Dniepr - Donetz depression (closed to the north-west), sandy deposition prevailed ("Kanev time"; MAKARENKO, 1987). It was separated by the Donetz and Ukrainian land masses from the major depositional areas in the south. Sandy, glauconitic and siliceous sediments accumulated in shallow-water shelf environments extending from the Southern Ukraine via the Volga - Don area towards the (Pre-)Ural. The interplay of important fluxes of terrigenous sediments, run-off features and the influence of boreal water masses led to a differentiation in overall sedimentation / environmental conditions between northern and southern areas. The composition of Ypresian radiolaria and diatom associations from the Middle Volga and Precaspian areas strongly indicates the existence of open marine connections with both arctic and tropical realms (KOZLOVA, 1993). More fine-grained sediments (muds with calcareous nannoplankton, rare planktonic foraminifera and siliceous microfossils; BENIAMOVSKII *et al.*, 1990) accumulated in the southern and eastern parts. The southern parts of the Scythian plate and most of the Precaspian depression were subject to deposition of

calcareous sequences. Siliceous to calcareous muds (10 - 150 m) with rich associations of calcareous nannoplankton and foraminifera, or with nummulite associations of mixed European and Mediterranean affinities developed in the Crimean and Precaspian areas (NAIDIN & BENIAMOVSKII, 1994). Deeper-water foraminiferal marls (up to a few hundred metres) were deposited in the central parts of the Indol - Kuban, Terek - Caspian and Precaspian depressions. These marls belong to the *Morozovella subbotinae, M. marginodentata* and *M. lensiformis* regional zones (SHUTSKAYA, 1970; BENIAMOVSKII *et al.*, 1990).

II.4.2.2.- West Siberian platform, Turan plate and Middle Asian depressions

Siliceous sedimentation prevailed in the boreal West Siberian basin, as expressed, for instance, by the diatomites of the Irbit Suite (up to 50 m). In the northern Turgaj Strait, glauconite sands with gravel interbeds were deposited; accumulation of sandy, clayey and calcareous sequences predominated in the southern part of the strait and in the northern Pre-Aral area (15 - 25 m of the Taup Suite). The involved shallow-water environments extended to the west, along the western slopes of the Mugodzhary High. Carbonate sedimentation was predominant in the outer shelf areas of the Turan Sea. The resultant successions yield rich calcareous nannoplankton and planktonic foraminiferal associations indicative of the Early Ypresian *Morozovella subbotinae, M. marginodentata* and *M. lensiformis* regional zones. Their dark interbeds mirror the effects of repeated episodes of stagnant water masses. Shallow-water, sandy limestones with nummulites were deposited in the Mangyshlak shoal areas (NAIDIN *et al.*, 1996). The Turan Sea extended into Middle Asia as far as the Tadjic, Fergana and Tarim depressions (east of mapped areas). In these depressions fine muds with diversified faunal associations (corals, echinoids, molluscs, foraminifera) accumulated. These associations are indicative of an overall transgressive marine setting, including depositional environments with normal salinities (DAVIDZON *et al.*, 1982).

II.4.2.3.- Black Sea depressions and Greater Caucasian - Kopet Dagh basin

This deep-water basin system was subdivided by the Andrusov and Shatsky Ridges into the Western Black Sea, the Eastern Black Sea and the Greater Caucasian domains (SCHERBA, 1993). In the central parts of the Black Sea depressions up to 300 - 500 m of clayey sediments were deposited during the (undifferentiated) Palaeocene / Eocene, as can be inferred from seismic surveys (TUGOLESOV *et al.*, 1985). Calcareous flysch sequences with volcaniclastics derived from the Iran - Lesser Caucasian volcanic arc accumulated in the Adzharo - Trialetic areas of the Eastern Black Sea domain (300 m of the Borzhomi Flysch with redeposited nummulites; PALAEOGENE SYSTEM, 1975). The Shatsky Ridge was a shoal area with deposition of nummulite limestones and marly sediments; no sedimentation occurred on the steep southern slope of the ridge. The eastern, Kura part of the former Mesozoic island arc subsided to bathyal depth (GEOLOGY OF THE USSR, 1964), associated with the accumulation of calcareous muds. The axial part of the Greater Caucasian basin was a deep, steep-walled trough (located approximately along the present-day southern slope of the Greater Caucasus; KOPP & SCHERBA, 1998), which had subsided to about 4000 m in the Palaeocene, as inferred from seismics (TUGOLESOV *et al.*, 1985) and micropalaeontological data (BENIAMOVSKII & SCHERBA, 1999). The trough shallowed in the course of the Early Eocene. The position of the continental slopes in the north was located in the southern parts of the Indol - Kuban and Terek - Caspian depressions. Gravity-sliding, incision of canyon systems and non-deposition were characteristic features during the Early Eocene, whereas Palaeocene deposition was marked by turbidites and olistostromes. The Greater Caucasian basin passed into the South Caspian depression, where 40 - 60 m of finely-bedded calcareous oozes were deposited in the deepest parts.

II.4.2.4.- Iran - Lesser Caucasian volcanic belt and interarc basins (not mapped)

The uplifted, but still submerged Lesser Caucasus domain belonged to the southern part of the former Mesozoic island arc. Volcaniclastic sediments accumulated in the western parts; predominantly shallow-water, detrital algal limestones (*Lithothamnium*) and marls developed in other parts of the region. The Erevan - Ordubad interarc depression in the south was subject to calcareous flysch sedimentation (Garni Suite). The lower part of the flysch succession yielded Late Palaeocene to Early Eocene age-diagnostic planktonics (*Morozovella marginodentata* and *Acarinina subspherica*; PALAEOGENE SYSTEM, 1975). Thick, relatively proximal flysch sequences (up to 1200 m) were deposited in the southern parts of the depression, whereas up to about 500 m of distal flysch successions were deposited in the north (SADOJAN, *in* SCHERBA, 1993), pointing to a southern provenance area of the clastics ("Iran volcanic high").

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi-Oman domains of the Arabian peninsula

The Early to Middle Ypresian sedimentary record portrays a regional marine transgression which resulted in the development of the oldest Tertiary carbonate platform of Arabia as part of a shelf bordering the eastern African plate margin. Overall transgressive deposition and regional subsidence allowed the accumulation of thick sequences of shallow-water carbonates, rich in foraminifera and molluscs (100 - 150 m in Oman; ROGER *et al.*, 1992; LE MÉTOUR *et al.*, 1995; about 100 m in Saudi Arabia; POWERS, 1968). The accumulation of the carbonates occurred in association with subordinate deposition of bioclastic shales (Umm er Radhuma Formation). Clastic influxes were of minor importance, except for the marginal shelf areas, i.e., along the eastern rim of the proto-Oman mountains (Jafnayn Formation; NOLAN *et al.*, 1990). Eoalpine compressional tectonic deformation, initiated during the Late Cretaceous, affected the Early to Middle Ypresian palaeogeography along the eastern margin of the Arabian platform. The tectonic activity included the reactivation of the Muthaymimah trough in a flexural basin setting and resulted in concomitant gravity-induced sedimentation (LE MÉTOUR *et al.*, 1992, 1995).

The south-eastern part of Arabia (Abat and Masirah troughs) corresponded to a passive margin, onto which hemipelagic and (subordinate) gravity-flow deposits accumulated (Abat and Sirab formations; PLATEL *et al.*, 1992a; WYNS *et al.*, 1992).

II.5.2.- Israel

In the Levantine part of the Eastern Mediterranean region (BENJAMINI, 1979, 1980, 1984; BUCHBINDER *et al.*, 1988; HATZOR *et al.*, 1994) the development of depositional environments in Eocene times was controlled by folding in the domain of the evolving Levantide chain (Syrian arc). The tectonic movements resulted in the origin of various subbasins, separated by partly emerged ridges. The Ypresian consists of up to 200 m of nummulite limestones and pelagic, foraminiferal oozes (chalks with chert) of the Adulam Formation. They reflect deposition in shallow-water environments which developed along SW-NE oriented anticlinal ridges, and sedimentation in basinal deep-water settings, respectively. The shallow-water carbonates are onlapping the islands which may have formed during periodical uplift of the Levantide chain. The abundance of slump structures corroborates the assumption of episodic tectonic instability.

II.5.3.- Egypt

The maximum transgression during the Cainozoic occurred during the Late Palaeocene, when marine depositional domains extended as far as northern Sudan. From the Eocene onward, the Cainozoic history was characterised by an overall retreat of the sea towards the north (SAID, 1990). Facies developments in the Early Tertiary basin systems of Egypt and neighbouring regions were strongly controlled by folding and faulting related to the evolution of the Syrian Arc system. Early to Middle Ypresian (P6 / NP10) sediments, conformably overlying the Upper Palaeocene, reflect deposition in open marine environments (green shales of the Esna Formation in Upper Egypt and the Sinai region; chalks and marly sediments in the west Farafra Oasis and north-eastern Sinai, respectively). Limestones with chert and minor shale interbeds (lower part of the Apollonia Formation) accumulated in elongate basins between E-W oriented highs (coastal area and Kattaniya - Cairo / Suez highs) in the northern parts of the Western Desert. Early to Middle Ypresian successions reach their maximum thickness in the stable shelf sequences of Upper Egypt.

II.5.4.- Tunisia

In Eastern Tunisia and in its adjoining offshore region (Hammamet Gulf), Early Eocene sediments are absent in various, relatively small-sized areas (BONNEFOUS & BISMUTH, 1982). Jointly, these areas may have represented an "archipelago" which was composed of small islands and/or submarine palaeoreliefs surrounded by deep-water environments. Larger-sized areas which were emerged during the Early Eocene existed in Central (Kasserine island) and Southern Tunisia. Along the northern margins of the emerged domain of southern Tunisia, calcareous lacustrine deposits accumulated. The Bouloufa Formation, a continental unit composed of conglomerates, red clays, lacustrine limestones and caliches (dated by continental molluscs; ABDELJAOUED, 1983, 1991), is partly correlative with this lacustrine sequence. In the Tellian domain, the Ypresian portrays a change from clayey towards calcareous sedimentation in increasingly deeper-water environments (*Morozovella subbotinae*, *M. formosa* and *M. aragonensis* Zones). South of Kasserine island, shelly limestones belonging to the lower part of the Metlaoui Formation (BUROLLET, 1956; BÉJI-SASSI, 1999) and dolomites accumulated in shallow-marine, semi-enclosed environments favouring the genesis of phosphates. Towards the east (Jebel Faid and Jebel Lessouda; "North-South Axis"), coeval phosphatic sequences were deposited, interbedded with gypsiferous horizons (MATMATI *et al.*, 1992), or associated with marls, clays, lumachelles and cherty dolomites rich in molluscs, vertebrate remains and coprolites (Metlaoui - Gafsa region).

In the Jebel Trozza region (near Kasserine island), the Early Ypresian transgression is recorded by fossiliferous conglomerates with clasts reworked from the Cretaceous substratum (EL GHALI, 1993). North-west of Kasserine island (Maktar - Sers region), the Palaeocene - Eocene transition is marked by a glauconitic level, which is overlain by marls (upper part of El Haria Formation; SAID, 1978; ZAIER *et al.*, 1998) of the *Morozovella subbotinae* Zone. The latter formation was deposited in upper bathyal to circalittoral environments (EL KAROUI-YAAKOUB, 1999). It is overlain by the calcareous and detritic Metlaoui Formation (*M. aragonensis* Zone), which was deposited in shallower marine environments. More or less similar littoral environments of deposition developed during the Early Ypresian in the Bargou region (northern part of Central Tunisia), following the deposition of deep-marine Palaeocene clays and a thin phosphatic and glauconitic horizon of polymict microconglomeratic limestones around the Palaeocene - Eocene transition (BEN ISMAIL-LATTRACHE & BOBIER, 1996). In the Serj region (northern part of Central Tunisia), the Early Ypresian is represented by marls (*M. subbotinae* Zone) and calcareous deposits rich in nummulites (*N. exilis, N. deserti, N, praecursor*). Also in the Jebel Nara region ("North-South Axis"), the Early Ypresian successions reflect deposition in circalittoral environments; shallowing induced the development of a carbonate platform in this region. South of the Bargou region, micritic limestones were deposited. Upwards, they pass into the bioclastic limestones (yielding Early Ypresian larger foraminifera) of the El Garia and Jbile region (RIGANE & BENJEMIA-FAKHFAKH, 1991; BEN ISMAIL-LATTRACHE & BOBIER, 1996). Lagoonal conditions, evidenced by sedimentary sequences including (partly) brecciated dolomites and gypsum (composing the lower part of the Metlaoui Formation), prevailed in the Jebel Sehib - Jellabia area (Gafsa basin; CHAABANI, 1995). Both the thickness and facies distribution of the Ypresian deposits were controlled by (inherited) E-W, NW-SE and NE-SW fracture systems, reactivated by tectonics and halokinesis (ZAIER *et al.* 1998). In the Gafsa basin, a NW-SE trending extensional stress regime existed during the Ypresian (BOUAZIZ *et al.*, 1998), whereas transtension controlled sedimentation in Central Tunisia (RIGANE & BENJEMIA-FAKHFAKH, 1991).

18.- LATE LUTETIAN (44 - 41 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The Late Lutetian ranges from about 44 to 41 Ma ago, which interval covers the upper part of the *Globorotalia kugleri / Morozovella aragonensis* (P11) and the larger part of the *M. lehneri* (P12) planktonic foraminiferal zones (Fig. 17.1). These zones are correlative to the upper part of the calcareous nannoplankton zone NP15 and the lower part of NP16 (ANDREYEVA-GRIGOROVICH, 1973). In the eastern parts of the northern Peri-Tethys platform (Crimean - Caucasian areas) the Late Lutetian is considered equivalent to the Keresta and lower Kuma horizons; for these parts of the Peri-Tethyan realm the mapping interval corresponds to the Keresta time-window. In terms of mammal biochronology the Late Lutetian roughly includes (parts of) the MP12 - MP13 zonal interval.

I.2.- Structural setting and kinematics

In the course of the Middle to Late Eocene, large-scale northward-directed latitudinal shifts of the African and Eurasian plates gradually came to an end (DERCOURT *et al.*, 1986, 1993), in concomitance with the inception of mechanical coupling of the African/Apulian and Eurasian plates. In the south-west, the still ongoing motions of the African/Apulian plate relative to Europe were accompanied by southward thrusting of the Pyrenean orogenic belt and by the accumulation of foreland-basin depositional sequences derived from the evolving chain into the Ebro basin. To the north, a narrow foredeep persisted, after an earlier episode of major Pyrenean orogenic activity. In the peri-Alpine - Carpathian domains oceanic subduction prevailed. In the east, an approximately W-E trending volcanic chain developed in response to the extrusion of andesites and basalts all along the Tethyan - Peri-Tethyan transition zone from the Rhodope massif in the west via the Pontian - Lesser Caucasian domains towards the Talysh - Elburz domain farther in the east. On the western part of the northern Peri-Tethys platform, major parts of the Cainozoic rift system (Rhine, Bresse and Limagne grabens) originated in the course of the Eocene, i.e. from about the Ypresian - Lutetian transition onward (ZIEGLER, 1990, 1994; PRODEHL *et al.*, 1995; SISSINGH, 1998). On the Iberian block the separation between the Duero and Tagus basins became more pronounced in response to the development of the Central Range. No fundamental tectonic reorganisation seems to have taken place on the southern Peri-Tethys platform in Early/Middle Ypresian to Late Lutetian time.

I.3.- Outlines of palaeogeography and palaeoenvironments

Trans-European marine corridors, in particular the major ones connecting the Arctic Ocean (via the Siberian basin and the Turgaj Strait) and the Atlantic Ocean (via the North Sea/Polish Lowland basin, Pripyat Strait and Dniepr - Donetz depression) with the Tethyan - Indo-Pacific realm still largely defined environmental conditions and faunal/floral exchanges after the Early to Middle Ypresian. However, the marine connection of the West Siberian basin with the Arctic Ocean was disrupted in the course of the Lutetian. The Pripyat Strait had reopened after its (temporary) closure in the Early Eocene. Predominantly shallow-marine, clastic sequences accumulated on the northern Peri-Tethys platform. In latest Lutetian ("Kuma") time, recurrent stagnation events affected the deeper basins of the eastern parts of this platform from southern Ukraine / Crimea towards the Kopet Dagh domains This was followed by the disrupture of marine connections with the Arctic Ocean in the latest Lutetian. This, in turn, caused in the West Siberian basin a change from hitherto prevailing open marine environments to semi-enclosed domains characterised by reduced salinities. In the south-west (Pyrenean and Aquitaine regions) the marine corridor connecting the Atlantic and

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the Tethys was closed and the distribution of carbonate platforms at either side of the Pyrenean basin became more restricted. However, the marine realm continued to be open towards the Bay of Biscay. In the evolving European rift system different types of terrestrial, predominantly lacustrine sequences accumulated. They portray the contemporary absence of nearby pronounced reliefs. The grabens intersecting the Massif Central were open towards the Paris basin, which had been subject to a change from Early Lutetian marine carbonate platform towards brackish - lacustrine depositional environments in the Late(est) Lutetian. In the areas straddling the Tethys - Peri-Tethys transition, i.e. in the peri-Alpine and peri-Carpathian basins, flysch deposition persisted in Late Lutetian time. On the southern platform calcareous sediments accumulated over large parts of Arabia. Towards the west, the Early Ypresian terrigenous-clastic sedimentation made place for the deposition of platform carbonates on the eastern parts of the northern margin of the African plate.

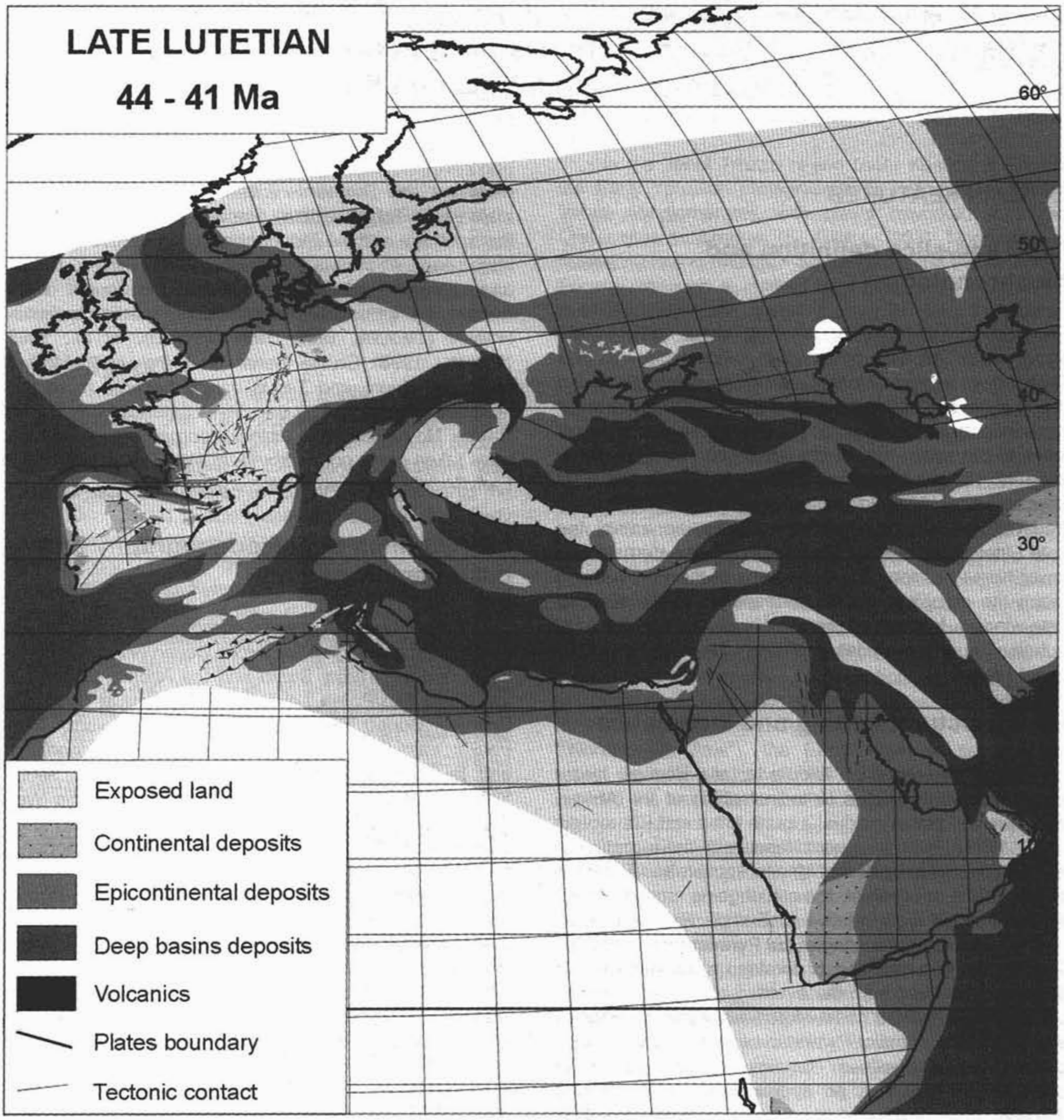

Fig. 18.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Late Lutetian.

II.- Description of Domains

II.1.- South-western Europe

II.1.1.- Iberian Peninsula

Deformation associated with the progressive narrowing of the Pyrenean basin resulted in the formation of imbricated thrust-sheets (Santanach, 1997). These

processes are mirrored by the foreland basin sequences of the Ebro basin. The emplacement of nappe units resulted in a further differentiation between sedimentary domains. The shortening of the foreland basin domains was coupled with a reduction of the extension of carbonate platforms, but the overall hierarchical organisation of the basins was basically the same as in the Ypresian. Siliciclastic deltaic depositional systems developed and fluviatile and alluvial sequences accumulated along the up-thrusted reliefs. A predominantly calcareous depositional area remained open to the Bay of Biscay. Predominantly alluvial sequences also accumulated along the margins of the large interior basins (Duero and Tagus basins). Both in the Tagus and Ebro basins evaporitic depositional systems were widely distributed. As in the Early Eocene, lacustrine carbonate sedimentation prevailed in the (central parts of the) Duero basin. The separation between the Tagus and Duero basins by the Central Range became more pronounced, in response to crustal thickening (VEGAS *et al.*, 1990). Displacements along major fault systems (Plasencia fault, Nazare - Lousa fault) in the central and western parts of the Iberian massif are thought to account for the basin evolution in these areas. The sequences of the peripheral depositional areas in Portugal are characterised by siliciclastic, continental sediments with well-developed pedogenic features (CUNHA, 1994), while lacustrine, calcareous successions were deposited during the Late Lutetian in the east, i.e. in the region of the later Valencia basin.

II.1.2.- Aquitaine basin

The marine depositional realm of the Aquitaine basin was reduced in size relative to the one during the Early to Middle Ypresian. The marine corridor towards the Tethyan realm was closed and (only) a narrow foredeep existed along the northern margin of the evolving Pyrenean orogenic chain. These changes in palaeogeographic configuration were induced during the major phase of the Pyrenean orogeny, which also resulted in uplift of anticlinal structures and diapirism of Keuper evaporites and in the origin of the North Pyrenean thrust front. In the south, Late Lutetian sedimentation is represented by the bathyal to circa-littoral "Marnes de la Côte des Basques" and "Marnes de Miretrain" formations, and by the littoral "Calcaires de Nousse" and "Calcaires de Brassempouy". The northern Aquitaine basin was characterised by the accumulation of neritic carbonates ("Calcaire de Blaye" *p.p.*, SZTRAKOS *et al.*, 1998). Rich and diversified larger foraminiferal associations with alveolines, nummulites and discocyclines indicate tropical climate conditions. Detrital sedimentation prevailed in the south-eastern basin. The resultant "Poudingue de Palassou" and "Poudingue de Jurançon" reflect the continuing supply of coarse clastics from the emerging Pyrenees and the filling-up of the eastern part of the Pyrenean basin. Also the north-eastern parts of the Aquitaine basin were subject to deposition of continental successions ("Formation de Guizengeard inférieure").

II.2.- Western Europe

II.2.1.- Paris basin

Following a brief episode of basin uplift and erosion around the Ypresian - Lutetian transition, deposition was continued during a major marine transgression which entered from the North Sea basin and affected the Paris basin throughout the Early and Middle Lutetian. Time-transgressively, sandy and glauconitic carbonates firstly accumulated in relatively high-energy, coastal marine environments with rich and diversified macrofaunas. Continued subsidence of the basin ultimately resulted into widespread deposition of limestones ("Calcaire grossier"), reflecting more open-marine conditions. During the Late Lutetian, less open-marine environments of deposition were established in the greater part of the Paris basin. Predominantly, marly strata were laid down. In the south, brackish-water lagoons prevailed. At that time, the basin was disconnected from the North Sea basin by the emerged Artois axis. A residual marine corridor existed to the (south)west of the Weald Anticline, allowing marine communication between the Paris basin and the Channel and Western Approaches basins (MÉGNIEN, 1980; GÉLY & LORENZ, 1991).

II.2.2.- Southern North Sea basin

Deposition in the North Sea basin during the Early Lutetian was characterised by the accumulation of glauconitic and calcareous sandstones with nummulites (main occurrence of Brussels Sand). Their deposition culminated during an Early Lutetian regression which may be related to a further emergence and erosion of the Artois Axis that overprinted eustatic transgressions. These sandy strata are overlain by comparable deposits (Lede Sand) laid down during a Middle Lutetian phase of dominantly eustatic transgression that was concluded by another phase of uplift of the clastic provenance area (Artois Axis) and the southernmost North Sea basin. During this tectonic phase, marine communication between the Paris basin and the North Sea basin was permanently interrupted. Thus, the (southern) North Sea basin developed boreal faunal characteristics from the palaeogeographical severance of these basins onwards. Tectonic uplift continued in Eastern Belgium until about the Priabonian - Rupelian transition. In north(western) Belgium, however, subsidence and resultant deposition was resumed towards the end of the Lutetian. In this part of the North Sea basin the successive, Late Lutetian phase of deposition was typified by transgressive sedimentation of sands and clays (Wemmel Sand, Asse Clay) which were conformably overlain by regressive strata (Ursel Clay). On their turn, the latter beds were succeeded by various sequences consisting of other, alternating transgressive and regressive glauconitic sands during the Late Lutetian - Priabonian (LETSCH & SISSINGH, 1983; VINKEN, 1988; VANDENBERGHE *et al.*, 1998).

II.2.3.- European rift system

The oldest, Lutetian graben-fill sediments are palaeogeographically unevenly distributed. In the Rhine graben they include the Messel Formation (MP11, up to 250 m), Eocene Basal Clay Formation (up to at least 60 m) and

Siderolithic Formation (over 100 m). These deposits accumulated during the early part of the Lutetian in lacustrine and terrestrial environments. The Late Lutetian is most notably represented by the Ubstadt - Bouxwiller beds (MP13, about 45 m), a sequence consisting of limestones, dolomitic marls, clays and lignites. These strata were deposited under warm and humid climate conditions in non-restricted lacustrine depocentres. Altogether, the Lutetian lithological complex of different continental sediments generally accumulated in palaeogeographically restricted depressions, lakes or river valleys. It reflects an initial phase of rifting that was typified by regionally differentiated subsidence related to strike-slip movements along pre-existing faults (SISSINGH, 1998). The paucity of coarse conglomerates indicates that the nearby palaeo-relief was probably modest, with the exception of the southern segment of the Rhine graben. The different sites of sedimentation received variable amounts of clastics. This probably depended mainly on the local rate of tectonic subsidence, which, in general, seems to have been in balance with the rate of sediment input. In the Hessen depression (Kassel basin), Lutetian clays, sands and lignites accumulated under more or less comparable conditions (MEIBURG & KAEVER, 1986).

In the Rhone graben (particularly so in its northern part, the Bresse graben) similar continental deposits accumulated from about the Ypresian - Lutetian transition onwards until the Bartonian - Priabonian transition (up to about 250 m; partly dated by the Late Lutetian / Bartonian to Priabonian mollusc *Dissostoma (sub)mumia*). Calcareous lacustrine deposits (yielding the Lutetian marker species *Planorbis pseudoammonius* in the Talmay Limestone) are more or less predominant. In general, the (Late) Lutetian Rhone graben corresponded palaeogeographically to a series of lakes (CAVELIER, 1984). The palaeorelief of the graben defined various localised lacustrine subbasins, sills and fluvial source areas. The lakes received clastics from low-lying landmasses along the eastern and western margins of the initial rift, as well as from intra-graben highs. In the course of the Lutetian and Bartonian, the lacustrine depositional environments became more widely distributed and more calcareous.

In the Massif Central rifting and deposition seems to have started at about the same time as in the Rhine and Rhone grabens. In the Limagne graben the initial basin fill comprises a fluvio-lacustrine sequence of sandy, marly and more calcareous sediments (up to 600-700 m; partly dated as Late Lutetian and Early Bartonian; RIVELINE *et al.*, 1988), whereas deposition of fluvial arkoses (60 m) predominated in the coeval Puy basin (REY, 1971; AUTRAN & PETERLONGO, 1980). During this first phase of syn-rift deposition, the surrounding Massif Central was still a uniformly pene-plained area of erosion and non-deposition. Volcanism was confined to a few centres. Fluvial drainage and transport of clastic material occurred probably from the Limagne graben towards the Paris basin.

II.2.4.- Alpine foreland basins

During the Lutetian, the Alpine foredeep continued to onlap the southern margins of the Alpine foreland in northern and western directions. In concomitance with flexural subsidence of the basin in front of the advancing Alpine orogenic wedge, the tripartite succession of marine limestones, marls and turbiditic sandstones continued to accumulate progressively at increasingly greater water depths in northern and western landward directions. Deposition of nummulite limestones persisted in the Bauges basin till mid-Lutetian time, when communication between this basin and the actual foredeep was temporarily interrupted. Following a Late Lutetian - Bartonian episode of renewed communication with the fore-deep and of carbonate deposition, the Bauges basin became permanently integrated into the deeper-marine foreland basin at the end of the Bartonian (KERCKHOVE, 1980; CAVELIER, 1984).

During the Priabonian, foredeep deposition was strongly controlled by tectonics and sea-level change. Prior to the accumulation of the typical Molasse series after the earliest Rupelian, the North Helvetic flysch was deposited in a "Pre-Molasse" basin, in which, next to these mass flow deposits, widespread shelfal limestones developed on top of partly fluvial sandstones in the eastern part of the basin (SISSINGH, 1997). Priabonian sediments accumulated predominantly during a transitional sedimentary cycle set under partly underfilled and partly steady-state foredeep conditions of deposition. Altogether, the sequence reflects sedimentary environments which range from deep-water marine to continental (fluvial) and which are characterised by the absence of classical molasse. Via the intermediary "Pre-Molasse" basin setting and under the continuing influence of the regionally evolving Alpine mountain belt and local syn-sedimentary faulting, the mainly deep-marine and underfilled North and West Alpine foredeep basins were ultimately transformed into the largely shallow-marine and continental, filled to ultimately overfilled North and West Alpine Molasse basins, which persisted during the Oligocene and most of the Miocene. These important transformations in palaeogeographic and depositional circumstances were induced in particular by a pronounced mechanical coupling of the converging Apulian and European plates. As a result of this event, thrust-loading of the European crust commenced on a major scale, in conjunction with intra-orogen uplift and erosion.

II.3.- Central Europe

II.3.1.- Polish Lowland basin

In the Polish Lowland basin a major transgression towards the east and south occurred during the Lutetian. Its sedimentary expression is characterised by the widespread distribution of fine-grained quartz - glauconite sands, ranging in thickness from 3 m (Siemen Formation) to 90 m (Szczecin Formation). Calcareous nannoplankton associations derived from the latter formation in the western area belong to the NP15 - NP16 zonal interval, while those recovered from areas in the north (lower Pomorze Formation) and south-east are indicative of NP 16 (ODRZYWOLSKA-BIENKOWA & POZARYSKA, 1978; GAZDZICKA, 1994). Benthonic foraminiferal associations with *Kolesnikovella muralis*, *Astacolus decorata*, *Pullenia quinqueloba* and *Heterolepa perlucida* are common in the Polish Lowland successions; similar associations are known from the Keresta horizon of the Crimea area. Radiolarians are abundant in the basal part of the

Szczecin Formation. Altogether, these microfossil associations allow to correlate the Polish Lowland sequences with those in Germany, Ukraine (Crimea) and Turkmenistan (ODRZYWOLSKA-BIENKOWA & POZARYSKA, 1978a).

II.3.2- Carpatho-Pannonian region *(not mapped)*

II.3.2.1. General features

The incipient Eastern Alps - Carpathian orogenic arc prograded probably more externally since the Early Eocene. Concomitantly, the bordering flysch basin systems narrowed. An archipelago developed behind the slightly curved external thrust front. The clastic sequences of the flysch basins were derived mainly from the emerged parts of surrounding platforms, intrabasinal sources and the developing mountain chain. The involved submarine fan systems were widely distributed. Along the eastern part of the active margin of the Alcapa microplate (Fig. 17.3), transgressive carbonate sequences accumulated either in fore-arc basins (Central Carpathian Palaeogene basin) or in an epicontinental, intra-plate setting (North Hungarian Palaeogene "Buda" basin).

II.3.2.2.- Regional aspects

In Lutetian times, the depositional area of the Outer Carpathian flysch changed. Deepening led to the deposition of predominantly variegated clays, thin-bedded turbidites and hieroglyphic beds relatively close to the North European platform (e.g., Subsilesian, Silesian, Dukla and Skole units; see Fig. 17.3). In the depositional domains of the Subsilesian and Skole units sediment transport by longitudinal currents prevailed; the clastics were derived from the platform, as well as from intrabasinal sources (e.g., the presumed "Baska Cordillera" located between the Silesian and Subsilesian domains). The sediment successions of the Silesian unit portray rhythmic deposition of pelitic, fine to medium and coarse-grained greywackes and arkosic sandstone turbidites. Most of the clastics were derived from the "Silesian Cordillera"; additional provenance areas were the Bohemian massif and uplifted parts of the incipient Carpathian mountain chain (RAKUS *et al.*, 1990). In the Dukla unit, located more to the east, predominantly distal turbidites and variegated clays were deposited; sediment transport (by longitudinal currents) was from the SE to the NW. In the Magura basin the deposition of fine-grained turbidites ended (first in the depositional areas of the southern Biele Karpaty and Krynica subunits) while the deposition of coarse-grained sandstone rhythmites increased. The occurrence of (rare) nummulites in the sandstones indicates transport from littoral areas to deeper-water environment (MISIK *et al.*, 1985). In general, the clastics were often derived from older flysch deposits of the evolving Outer Carpathian accretionary wedge. The evolution of the accretionary prism of the active Carpathian margin was associated with a coeval shift in position of the longitudinal axis of the asymmetric flysch basins towards the platform (POPRAWA & NEMÈOK, 1989). Both the "Silesian Cordillera" in the north and the "Magura Cordillera" in the south-east acted as provenance area of clastics (RAKUS *et al.*, 1990).

Large parts of the domain of the western Central Carpathians situated on the Alcapa microplate were invaded by the sea during the Lutetian. The transgression extended far land-inward; probably a connection existed between the Central Carpathian Palaeogene (fore-arc) basin and the North Hungarian Palaeogene (epicontinental) "Buda" basin (ANDRUSOV, 1965). Uplift affected only the internal zones of the central Western Carpathians (KOVAC *et al.*, 1994a). In the basins of the eastern segment of the Central Carpathians (Maramures area), limestones and sandstones were deposited. The successions of the Central Carpathian Palaeogene basin (transgressive sequence of the basal Borove Formation; GROSS *et al.*, 1984) are mainly composed of breccias, conglomerates, sandstones with nummulites, siltstones, clays and limestones; upwards they pass into Late Eocene deep-water clays and turbidites. Sedimentation in the epicontinental domain of the Transdanubian Central Range started with deposition of the coal-bearing Darvasto and Dorog formations in limnic to paralic environments. This was followed by the deposition of shallow-marine nummulite limestones, which, in turn, pass upwards into late Middle Eocene pelagic marls (Padrag Formation) deposited in bathyal environments (BALDI & BALDI-BEKE, 1985; BALDI-BEKE & BALDI, 1991; CSASZAR, 1997). Predominantly turbiditic sands and grey clays accumulated in the (marine) Szolnok flysch basin situated at the northern margin of the Alcapa microplate (NAGYMAROSY & BALDI-BEKE, 1993; NAGYMAROSY, 1998).

The structural position and overall configuration of the Late Lutetian sedimentary basins in the south-eastern Outer Carpathians were roughly similar to those existing during the Early Eocene, but the longitudinal axis of the flysch basin system may have shifted towards a more external (eastern) position. This implies that the zone of interfingering of clastics from western (archipelago) and eastern (platform) sources was equally re-located more to the east than during the Ypresian. Sedimentation rates were higher in the western than in the eastern parts of the basin system. The sedimentary records indicate great differences in facies development. These differences reflect a complex basin evolution, which was probably related to tectonics to the west of the basin (BADESCU, 1998) or to intra-basinal tectonic activity (SANDULESCU, 1992). Increased sedimentary loading in the west caused increased flexural bending of the platform and a concomitant basin-inward shift of provenance areas. In the external flysch zone of the Eastern Carpathians calcareous sediments (Paszieczna and Doamna limestones) accumulated in addition to sands and clays. The depths of the Carpathian trough ranged up to 1000 m (BADESCU, 1998); its water masses were well-oxygenated. Deviating salinities occurred only in the shallow(er)-water environments of the "Transylvanian Gulf", in which gypsum and dolomite were deposited (POPESCU, 1976). The composition of the Late Lutetian faunas indicates warm climate conditions also in these parts of the Carpatho-Pannonian domain.

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

During the Late Lutetian, the southern part of the Mesozoic island arc became subject to intensive andesitic - basaltic volcanic activity. The activity of the Rhodope -

Talysh - Afghanistan volcanic belt led to the separation of the Black Sea depressions and the Greater Caucasian basin from the Tethyan realm (KAZMIN *et al.*, 1987). The marine connections between the North Sea basin and the epicontinental basins of the eastern parts of the northern Peri-Tethys platform were re-established in Late Lutetian time. This can be inferred from, for instance, similarities in the composition of benthic foraminiferal (nummulites), calcareous nannoplankton and siliceous (radiolaria, diatoms) associations. The Ukrainian shelf domains and the Donetz massif had been transformed into an archipelago. Recurrent stagnation events affected the deep-water environments from the southern Ukraine / Crimea area towards the Kopet Dagh, particularly so in the latest Lutetian ("Kuma time"). In the course of the Lutetian, the marine corridor towards the Arctic Ocean was closed, resulting in the disappearance of open marine conditions and the establishment of depositional environments with reduced salinities in a semi-enclosed setting. These palaeogeographic changes were accompanied by the disappearance of siliceous planktonics (radiolaria, diatoms) and by the inception of predominantly clayey sedimentation. Elsewhere, from northern Ukraine in the west towards the northern Precaspian, North Ustjurt and the Pre-Tien Shan areas in the east, sand and siltstone successions (with dinoflagellates and siliceous planktonics) were deposited in shallow-water environments. The southern outer shelf zone was subject to accumulation of calcareous sediments. These sediments contain rich and diversified associations of planktonic and benthic foraminifera and calcareous nannoplankton. The occurrence of *Hantkenina* and *Globigerinatheka* representatives is indicative of a warm, tropical climate during the Late Lutetian (corresponding to the warmest interval of the Cainozoic). The existence of a temperature zonation at the time is suggested by the persistent presence of tropical taxa in the Pre-Caucasus associations, the exclusive occurrence of *Globigerinatheka* in those of the southern Voronezh and the absence of both *Globigerinatheka* and *Hantkenina* in the Kiev marls of the Dniepr - Donetz depression. Palaeobotanical evidence points to warm and arid climate conditions during the earlier Late Lutetian ("Keresta time"), whereas both floral and planktonic foraminiferal data suggest a shift towards a cooler and more humid climate during the latest Lutetian "Kuma time".

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

As earlier in the Eocene, the Russian landmass constituted the main source of clastics deposited in the eastern part of the northern Peri-Tethys platform. The Syrt and Ural highs were the principal positive topographic features of the landmass. Shallow-water conditions prevailed in the Dniepr - Donetz basin. The southern parts of the East European platform and all domains of the Scythian plate formed part of the vast "Scythian Sea". The coarsest sediments (sands and siltstones) were deposited in the Pripyat and Dniepr - Donetz depressions, along the southern margins of the Voronezh anticlinal zone and in the eastern part of the Precaspian region. These clastic sequences yielded rich and diversified benthic foraminiferal associations, calcareous nannoplankton, siliceous plankton and, rarely, planktonic foraminifera. Relatively clayey deposits with abundant siliceous microfossils (e.g., KOZLOVA *et al.*, 1998) accumulated in the northern and eastern parts of the Precaspian depression. The latter associations corroborate the presumed exchange of water masses with the North Sea basin and the Atlantic. In "Keresta time", foraminiferal / calcareous nannoplankton oozes were deposited in the outer shelf zones. This type of facies was widespread in the central part of the Dniepr - Donetz depression and in the depositional areas of the southern Ukraine, the Steppe-Crimea, the Fore-Caucasian and the south-western Precaspian domain. Rare bentonite intercalations are correlative with volcanic eruptions farther to the south. In "Kuma time", the Greater Caucasian - Kopet Dagh basin was separated from the Tethyan realm, which change was associated with the development of stagnant bottom waters.

II.4.2.2.- West Siberian platform and Turanian plate

The vast West Siberian basin had lost its connection with the Arctic Ocean in the Middle Lutetian (AKHMETIEV, 1995) and became thus transformed into a semi-enclosed basin with reduced salinities (BENIAMOVSKII *et al.*, 1999). Biogenic, siliceous sedimentation was replaced by the accumulation of clays, associated with the disappearance of polyhaline siliceous and calcareous planktonic associations (BENIAMOVSKII *et al.*, 1993). At the time, dinoflagellates became predominant in the West Siberian basin and in the Turgaj area (VASILEVA, 1990). Similar shallow-marine, clastic sedimentation and predominance of dinoflagellate cysts occurred all over the northern and eastern Turan areas (North Ustjurt, North Aral, South Turgaj and Pre-Tien Shan). Carbonate deposition was characteristic for the central and southern parts of the Turan Sea (SOLUN, 1975) and for the Tadjic and Fergana basins (east of mapped area).

II.4.2.3.- Black Sea depressions and Greater Caucasian - South Caspian basin

The deeper-water environments of the Black Sea - Greater Caucasus depositional realm were subdivided by the Andrusov and Shatsky ridges into three major depressions. Marly to clayey deposits (100 - 300 m) were laid down in the deepest parts of the Greater Caucasian trough. These deposits pass into (up to 1000 m thick) calcareous flysch sequences towards the marginal areas to the south. The Shatsky Ridge formed a shoal with (partly detrital) nummulite limestone sedimentation in its eastern part (Georgia). No coeval sediments accumulated on the steep, southern slopes of the ridge (TUGOLESOV *et al.*, 1985). The Eastern Black Sea depression extended into the Adzharo - Trialet depression, which was separated from the Talysh depression. These depressions were subject to andesitic - basaltic volcanic activity (KAZMIN *et al.*, 1987); sedimentation was characterised by the accumulation of thick (up to 4000 m in the Tbilisi area) flysch sequences (KAZMIN *et al.*, 1987). The shallow-water marginal shelf environments bordering the Pontides - Lesser Caucasus island arc (characterised by abundant

larger foraminifera including *Nummulites laevigatus* and *Discocyclina*) regressed to the north in the course of the Lutetian (GEOLOGY OF USSR, 1964).

II.4.2.4.- Iran - Lesser Caucasian volcanic belt and inter-arc basins (not mapped)

Shallow-water carbonates and volcaniclastics (up to 1500 m) accumulated on the slopes of the Rhodope - Talysh - Afghanistan volcanic belt. Thick flysch sequences were deposited in the Erevan - Ordubad inter-arc depression. The Somkhet - Akdam volcanic ridge was the main provenance area of these clastics. The relatively fine-grained, partly calcareous flysch successions deposited in the Araks area were derived from the South Armenian High (SADOJAN *in* KOPP & SCHERBA, 1998).

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi-Oman domains of the Arabian peninsula

A Late Lutetian rise in sea level led to a new major flooding of the Arabian platform and, consequently, to a new phase of development of widely-extended carbonate platforms. Regional subsidence resulted in the deposition of about 50 - 100 m of carbonates in Oman (ROGER *et al.*, 1992; LE MÉTOUR *et al.*, 1995) and in Saudi Arabia (POWERS, 1968). In contrast, several hundreds of metres of sediment accumulated in the concurrently subsiding (residual) Muthayminah trough (BÉCHENNEC *et al.*, 1992). The Late Lutetian sediments deposited in inner shelf domains mainly consist of foraminiferal carbonates, associated with bioclastic marls and shales (Dammam Formation), whereas those developed in the outer shelf areas in the region of the northern present-day Oman mountains (Muthaymimah trough) are composed of shales (LE MÉTOUR *et al.*, 1992). Accumulation of (reworked) clastics (Seeb Formation) in marginal shelf environments evidence the emersion of the eastern margin of the proto-Oman mountains in (Late) Lutetian times (NOLAN *et al.*, 1990; WYNS *et al.*, 1992). In the Abat trough (south-eastern margin of the Oman mountains) sedimentation was controlled by synsedimentary tectonic deformation induced by a SW-NE oriented extensional stress regime. As a result, very thick calcareous and siliciclastic sequences and gravity-flow deposits were laid down in shallow-water shelf environments; they include clay interbeds deposited in deeper water (Musawa Formation; WYNS *et al.*, 1992; LE MÉTOUR *et al.*, 1995).

II.5.2.- Israel

Differential vertical motions related to folding of the Levantid chain persisted during the Lutetian. The palaeogeographic configuration of SW-NE oriented subbasins separated by anticlinal structures, continued to control sedimentation (BUCHBINDER *et al.*, 1988). This can be inferred from the characteristics and distribution of the nummulite limestones of the Matred Formation (40 m) and the pelagic chalks of the Maresha Formation (100 m).

II.5.3.- Egypt

The overall regression towards the north, which had started after the Late Palaeocene, continued in the Late Lutetian. Predominantly shallow-water shelf sediments were deposited (mainly nummulite limestones with marly and shaly interbeds). Due to uplift and erosion along the northern highs, Eocene deposits are absent in the northern Sinai and in some off-shore areas (JENKINS, 1990). Eocene sediments were not penetrated or proved to be absent in most wells drilled in the Nile delta region (HARMS & WRAY, 1990).

II.5.4.- Tunisia

Due to erosion associated with uplift related to diapirism, Late Lutetian sediments are absent in a WSW-ENE trending zone from Jebel Hairech to Protville (Bizerte). However, adjacent to this zone, well-exposed marine sediments of this age occur. Other "shoals" devoid of Late Lutetian sediments were encountered in wells drilled in Eastern Tunisia and in the adjacent offshore area, as well as in the Sahel region. In central-west Tunisia the Kasserine island still existed. Also the region of Southern Tunisia was emerged. It was separated from the Kasserine island by the Gafsa basin in which very shallow depositional environments with fluctuating salinities were predominant. Adjacent to and locally on the Kasserine island, lacustrine reddish and greenish clays and siltstones were deposited (TRUC, 1981, 1989; SASSI *et al.*, 1984). Lacustrine deposits (rich in carbonate concretions and well-cemented limestones) also accumulated along the northern margins of the emerged, non-depositional domain of Southern Tunisia (Bouloufa Formation; ABDELJAOUED, 1983). Palaeosoils and fluviatile sandstones developed in the Hachichina region (ABDELJAOUED, 1983; ABDELJAOUED *et al.*, 1984). In the western and central parts of Tunisia (Kef, Tajerouine and Tala regions) limestones and marls rich in pelecypods, echinoids and larger foraminifera (*Nummulites gizehensis*) were deposited. In the Jebel Nara and Jebel Cherahil regions a thick sequence (240 - 460 m) of marls with numerous lumachelle interbeds and, locally, gypsum accumulated contemporaneously in shallow-marine environments with fluctuating salinities ("Coquina facies" of the Cherahil Formation; COMTE & DUFAURE, 1973), as indicated by the benthic faunal associations (pelecypods, echinoids) and the absence of planktonic foraminifera (CASTANY, 1951; MATMATI *et al.*, 1992). West of Kasserine island, these successions contain gypsum layers, indicating a transition towards hypersaline conditions.

In the Tellian domains of northernmost Tunisia, marls and clays with rich planktonic and benthic foraminiferal associations (ALOUANI *et al.*, 1996; ROUVIER, 1977) accumulated in deep-marine environments. Shallow-water, nummulite limestones were locally deposited along the margins of the Hairech - Teboursouk shoal in the north (ROUVIER, 1977). To the south of this shoal and along the north-eastern margin of the Kasserine island, conglomerates and dolomitic limestones with fine-grained clastic intercalations overlain by lumachelles were laid down in shallow-marine environments (COMTE & DUFAURE, 1973). Lumachelles developed also in the Trozza region. Deeper-marine clays with rich and diversified foraminiferal associations accumulated in

north-eastern Tunisia (Souar Formation; BUROLLET, 1956; BEN ISMAIL-LATTRACHE & BOBIER, 1984); an intercalated level of upper bathyal to littoral clayey limestones (Reneiche limestone) separates the lower and upper members of the Souar Formation. In the north-west of its distribution area, the Reneiche level contains numerous larger foraminifera, including age-diagnostic nummulites (BLONDEAU, 1970, 1980). Open marine, relatively deep-water marly sediments with some limestone intercalations also accumulated in the Bargou region; they belong to the (regional) Lutetian ostracode *Loculicythere semipunctata* zone (BISMUTH *et al.*, 1978). In the Jebel region, coeval deposits including lumachelles and larger foraminiferal limestones represent the shallow-water equivalent of the deeper-water successions of the Bargou region. The present-day offshore regions of the Hammamet Gulf and the Pelagian Sea were characterised by platform carbonate sedimentation (including the Halk el Menzel Formation; BISMUTH & BONNEFOUS, 1982), as evidenced by exploration drilling and exposures on the island of Lampione in the Pelagian Sea. South-east of Kasserine island, thick evaporite (gypsum) sequences accumulated, along with some reddish and greenish clays and limestones indicating hypersaline, semi-enclosed environments (BUROLLET, 1956; BEJI-SASSI, 1985, 1999). Time-equivalent evaporitic successions were laid down in the Metlaoui basin, south of Kasserine island (CHAABANI, 1995).

19.- LATE RUPELIAN (32 - 29 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The Late Rupelian includes the calcareous nannoplankton zone NP23 and the early part of NP24, which interval is roughly equivalent to the planktonic foraminiferal zones P19, P20 and the earliest part of NP21 (Fig. 17.1, see also ANDREYEVA-GRIGOROVICH, 1981; NAGYMAROSY & VORONINA, 1993). The corresponding numerical ages for the lower and upper interval limits are about 32 and 29 Ma. The mapped interval for the Central Paratethys includes the upper part of the Kiscellian stage. In the Eastern Paratethys a slightly more confined interval, corresponding to the Solenovian regional stage, has been selected. Mammal associations recovered from the Late Rupelian interval cover (parts of) zones MP22 to ?MP24.

I.2.- Structural setting and kinematics

The effects of African/Apulian - Eurasian continent-continent collision became more pronounced in the Late Eocene (Priabonian). This resulted in a fundamental reorganisation of both the Tethyan and Peri-Tethyan realms, as evidenced, for instance, by the uplift and subsequent emergence of (parts of) the evolving Alpine chains from the Pyrenees in the west to the Caucasus and Elburz in the east. The processes in the Tethyan domains were coupled with large-scale uplift of both the northern and southern Peri-Tethys platforms. On the northern Peri-Tethys platform a Late Eocene (Priabonian) and an earliest Oligocene (Early Rupelian) phase of rifting resulted in the "maturation" of the European rift system.

I.3.- Outlines of palaeogeography and palaeoenvironments

With respect to overall palaeogeographic configurations, the Middle to Late Eocene phase of continent - continent collision resulted in the break-up of the Tethyan realm into various southern, "circum-Mediterranean" and northern domains by the end of the Eocene. From the Eocene - Oligocene transition onward the northern domains became subject to recurrent isolation from the Mediterranean and the Indo-Pacific through the initial uplift and emergence of the Alpine - Carpathian chain, the Dinarides, Balkanides, Hellenides, Pontides and parts of the evolving Lesser Caucasian and Elburz orogenic belts and the Kopet Dagh area. These domains are referred to as the "Paratethys". The Paratethyan realm became subdivided into the Western, Central and Eastern Paratethys, corresponding approximately to the peri-Alpine foreland basin domains, the depositional areas straddling the collision zone in Central Europe, and their equivalents in Eastern Europe and Western Asia, respectively. In the Late Rupelian, the marine connections of the Paratethys with Mediterranean basins were severed and the Paratethys had become completely isolated from the Indo-Pacific. Marine connections to circum-Mediterranean basins were most probably confined to a narrow corridor adjacent to the emerging Western Alps (RÖGL, 1998). To the north (North Sea basin) a marine seaway existed via the Rhine graben and Hessen depression. Another, although not open-marine corridor between Mediterranean and Paratethys might be hypothesized across the present-day Aegean region. However, the isolation of the Paratethys from the world's oceans was more severe than in any other episode during the Oligocene and Early Miocene. Intra-Paratethyan connections existed throughout the Late Rupelian, as follows from faunal evidence, and there is a remarkable correspondence in the main characteristics of sedimentation and environmental conditions from west to east. These conditions were characterised by reduced salinities all over the Paratethyan realm for at least part of the Late Rupelian and by the widespread occurrence of dysoxic or anoxic conditions, often associated with the accumulation of organic-rich muds and "menilites" in deep basins.

The seaway across the European platform towards the West Siberian basin via the Turgaj Strait was closed already by the end of the Eocene; that from the North Sea basin towards the Eastern Paratethys (via the Polish

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Lowland basin and Pripyat Strait) was closed in the course of the Rupelian. Similarly, marine environments had almost entirely ceased to exist on the Iberian block. Sedimentation in the Ebro basin changed into the accumulation of exclusively non-marine clastics, which were mainly derived from the emerging Pyrenees. In contrast, the Late Rupelian successions of the Aquitaine basin witness of a major transgression. On the southern Peri-Tethys platform the impact of continent - continent collision caused the uplift and emergence of almost the entire Arabian area. The widespread andesitic volcanism in the north-eastern parts of the African domain and in south-western Arabia may be considered the prelude to subsequent rifting resulting in the break-up of the African / Arabian block and the development of the Red Sea - Gulf of Aden rift system.

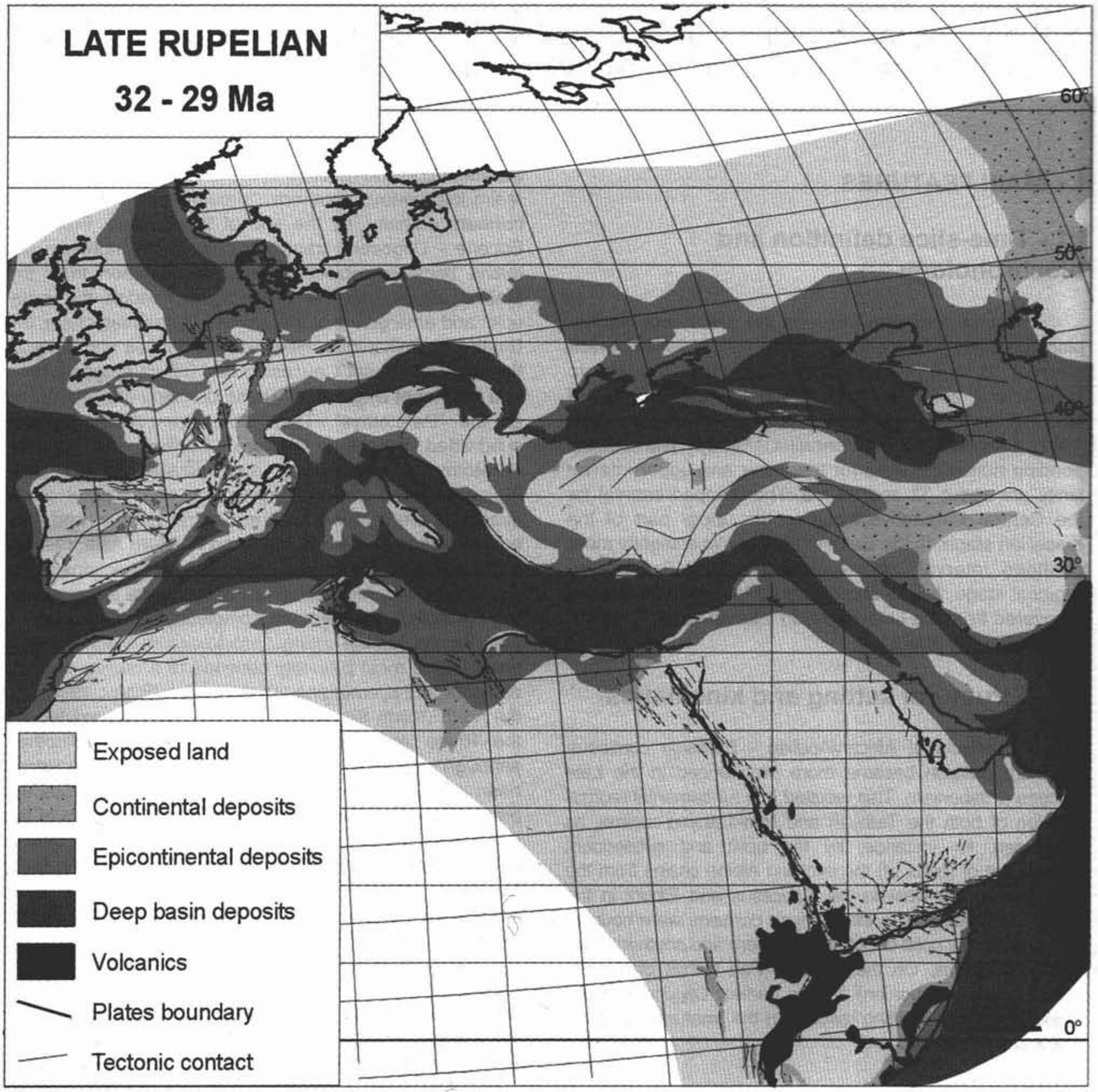

Fig. 19.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Late Rupelian.

II.- Description of Domains

II.1.- South-Western Europe

II.1.1.- Iberian Peninsula

N-S and NW-SE compression continued into the Oligocene and resulted in the further, southward progradation of Pyrenean thrust sheets and the foreland basin (Ebro basin) system (Sanz de Galdeano, 1996; Santanach, 1997). In the Cantabrian and Bay of Biscay domains compressional deformation processes had

decreased relative to the Eocene. The compressional regime in the northern parts of Iberia was approximately coeval with extension farther to the east, in the domains of the present-day (north)Western Mediterranean (SANZ DE GALDEANO, 1996). Due to the combined effects of extension in the domains of the Atlantic and Western Mediterranean, the eastern sector of Iberia was subject to compression. This, in turn, resulted in the westward movement of the Altomira Range in the Tagus basin region, as well as in the reactivation of Late Hercynian faults (e.g., transcurrent, sinistral displacements along the Plasencia fault and movements along faults delimiting the Central Range system). The latter system, separating the Duero and Tagus basins, became a distinct topographic feature in the Early Oligocene. No marine sediments of any importance accumulated in the Iberian domains in the Late Rupelian. Alluvial and lacustrine sedimentation was predominant in the large interior basins. Lacustrine sedimentation in the Ebro and Tagus basins was characterised by the accumulation of evaporites; deposition of carbonates occurred in the Duero basin. Some depositional areas with continental sedimentation developed in the domain of the Betics; similarly, continental deposits were laid down in some small interior and peripheral basins of the Iberian peninsula.

II.1.2.- Aquitaine basin

A major transgression occurred in Middle to Late Rupelian times, associated with the accumulation of neritic, calcareous sequences. The latter mainly correspond to the "Calcaire à Astéries" (particularly well-developed in the northern part of the basin; see PRATVIEL, 1972; GAYET, 1985). In the south-westernmost area, deposition of the Biarritz Sandstone Formation prevailed (MATHELIN, 1988). The rich and diversified faunal / floral associations of this age indicate tropical climate conditions (CAHUZAC, 1980). This inference is corroborated by the local presence of coral reefs. Sedimentation of pelagic marls occurred in the westernmost part of the basin (including the present-day offshore area). Molasse-type sediments accumulated in the eastern parts of the basin. Locally, lagoonal and lacustrine deposits were laid down. Characteristic formations are the "Sables du Périgord" in the north-east and the "Molasse de l'Agenais" in the east to south-east. Coarse clastics (conglomerates) accumulated in the Aquitaine basin along anticlinal structures and diapirs in the west and south, trending parallel to the WNW-ESE oriented Pyrenean orogen.

II.2.- Western Europe

II.2.1.- Paris basin

In Bartonian (Auversian - Marinesian) to Priabonian (Ludian) times, deposition continued with several short-term sedimentary cycles (GÉLY & LORENZ, 1991). Accumulation of sediment was interrupted by brief episodes of basin wide emergence and erosion / non-deposition. The corresponding hiatuses can be traced throughout the Paris basin. Lagoonal and shallow-marine environments of deposition prevailed in the basin proper. Continental strata were laid down along the southern basin margins. They became increasingly widespread during the Bartonian, while the basin decreased in size. At the beginning of the Priabonian, a major marine transgression from the north-west occurred. In response to a later Priabonian rise in sea level, the Artois Axis was again flooded, but without renewal of marine communication with the North Sea basin. Overall, the transgressions re-established a large continental-lagoonal, partly marine realm of deposition in the region of the Paris basin. Continental evaporites accumulated in the basin centre. Lagoonal conditions of deposition predominated in the Early Rupelian, another period of transgressive deposition (GÉLY & LORENZ, 1991). Following a brief mid-Rupelian period of tectonics-related, basin-wide emergence (Etrechy erosive phase), deposition in shallow-marine environments became more pronounced during a successive Late Rupelian rise in sea level (represented by the Fontainebleau Sands). The successive Chattian - Aquitanian depositional cycle is typified by the development of lagoonal marls and lacustrine limestones ("Calcaire de Beauce") which signify the demise of the Paris basin (GÉLY & LORENZ, 1991; MÉGNIEN, 1980). In concomitance with uplift, relief steepening and increased erosion of the Massif Central from the Aquitanian - Burdigalian transition onwards, the region of the previous Paris basin was encompassed into a large area with localised deposition of fluvial sediments (Orléans and Sologne sands). These Neogene clastics were distributed by northwards flowing rivers draining the elevated Massif Central, most of all involving the Palaeo-Cher and -Loire, and the Pre-Seine.

II.2.2.- Southern North Sea basin

Following a brief episode of tectonic uplift straddling the Priabonian-Rupelian transition, deposition of clastics was resumed in the southern part of the North Sea basin in response to subsidence induced by a first-order eustatic transgression that affected all of this depositional realm. Subsequent to the latest Priabonian accumulation of shallow-marine to lagoonal and terrestrial sands and clays with lignites (Klimmen, Grimmertingen and Neerrepen sands, Goudsberg clays) and some notable sandstones at the base of the Rupelian succession (Ruisbroek and Berg sands), clays with septaria (Boom Clay) accumulated all over the basin under generally open marine, relatively offshore shelf conditions. Deposition of these clays persisted until the Rupelian - Chattian transition when a global marine regression occurred in association with Late Rupelian - Early Chattian tectonic uplift of the bordering Artois axis (Ardennes - Brabant area). This regression resulted in basin-wide deposition of open-marine clays (Veldhoven Clay) which interfinger to the south with prograding coastal marine sands (Voort Sand). Much of this Chattian sequence was subsequently eroded in response to tectonic uplift and concomitant regression during the Aquitanian (LETSCH & SISSINGH, 1983; VINKEN, 1988; VANDENBERGHE *et al.*, 1998).

II.2.3.- Alpine foreland basins

The initiation of the post-Priabonian phase of deposition in the North and West Alpine Molasse basins coincided with a major eustatic rise in sea level during the earliest Rupelian. In the northern basin, the deep-water Fish and Engi shales were deposited. Their deposition was accompanied by rapid flexural subsidence of the

basin. Part of the basin was badly ventilated during the Early Rupelian, as testified by the deposition of anoxic shales. Deposition of shales occurred simultaneously with shedding of turbiditic clastics derived from the evolving Alpine orogenic wedge. Prior to further, Late Rupelian transgressive depositional onlap, a regression occurred, probably in response to both a eustatic fall in sea-level and tectonic uplift of the basin (SISSINGH, 1997). Successively, a major glacio-eustatic regression affected the Alpine Molasse basin at about the Rupelian - Chattian transition (LEMCKE, 1983). This general fall in sea-level was overprinted by short-term relative rises in sea-level, which are thought to have been forced by thrust-loaded subsidence of the Alpine foreland (DIEM, 1986). From that mid-Oligocene event onwards, non-forced regressions were responsible for the accumulation of predominantly continental deposits in the central and western parts of the North Alpine Molasse basin until the end of the Aquitanian (BERGER, 1996; SISSINGH, 1997). In the orogen-proximal zone of the eastern part of the basin, however, deep-water marine and flysch deposition persisted throughout the Chattian - Aquitanian.

During the Late Rupelian, a shallow, wave-dominated seaway with apparently fluctuating salinity and sandy tropical beaches and paludal shores seems to have been skirting the northern Alpine orogenic front. This elongate basin was drained axially from the west to the east. Concurrently, NW-SE oriented clastic shorelines with deltaic fans migrated eastward in the Swiss region during the terminal Rupelian. There is no evidence for Rupelian tidal activity (DIEM, 1986; HOMEWOOD, 1986). Following the mid-Oligocene regression, continental deposition prevailed in much of the northern basin. Thick, mainly conglomeratic alluvial-fan deposits accumulated. Northward flowing, marginal intra-Alpine streams formed a series of individual, sometimes laterally interfingering deltaic fans along the Alpine thrust front. The thrust wedge-fringing fans were characterised by steep proximal areas dominated by braided rivers. Down-stream, they gave way to relatively low-gradient floodplains typified by meandering rivers. Farther away from the rising mountain belt, lakes and swamps occurred. During this Chattian episode of deposition, a predominantly radial drainage system existed. However, with the increase of clastic input, attributed to rapid uplift of the Alpine orogen, this system deviated to the east and thus evolved into a largely longitudinally, east-directed drainage system (BÜCHI & SCHLANKE, 1977; BERGER, 1996; SISSINGH, 1997). This younger, Aquitanian system of meandering rivers was the first to cut a significant pattern of longitudinal valleys in the peri-alpine region. Thus the Rupelian - Aquitanian sequences represent a first-order upwards-shallowing and - coarsening cycle set, which commenced with a widespread phase of marine transgression and basin deepening and ended with a non-forced regression giving rise to predominantly continental sedimentation, once the volume of supplied sediment was equal to, or in excess of the accommodation space.

The North Alpine Molasse basin was continuous with the West Alpine Molasse basin, a shallow-marine gulf which was more or less curved around the western Alpine thrust front. The latter basin remained a marine depocentre till the end of the Rupelian when it was transformed into a fully continental realm of deposition. The closure of the marine basin corresponded thus in time to the glacio-eustatic fall in sea-level and the concomitantly large-scale, rapidly eastward-progressing continentalisation of the North Alpine Molasse basin around the Rupelian - Chattian transition. In the West Alpine Molasse basin the corresponding mid-Oligocene unconformity is also partly syn-tectonic (EVANS & MAGE-RAJETZKY, 1991). Apparently in response to progressing Alpine tectonics affecting the West Alpine foreland domain, continental strata accumulated in variably individualised areas of syn-tectonic deposition in a complex of interrelated alluvial facies (such as fans and ephemeral channels). The mid-Oligocene tectonic event, which uplifted and destroyed the marine West Alpine Molasse basin, resulted ultimately in the development of an erosional continental palaeomorphology with substantive differences in palaeorelief. Prior to the permanent closure of the basin, a northward-directed fluvial system that drained the continental region bordering the West Alpine orogenic wedge, most probably existed during the Chattian, in association with main directions of clastic supply directed to the (south)west and away from the Alpine fold-and-thrust belt.

II.2.4.- European rift system

The Middle Eocene episode of continental deposition in the initial Rhine graben was concluded by the development of a widespread hiatus that may be attributed to a Bartonian period of non-deposition and erosion that seems to have been induced by basin uplift in response to N-S directed compressional tectonic stress originating from the Alpine collision zone in the south (SISSINGH, 1998). A successive main phase of northward-progressing rifting induced pronounced subsidence in Priabonian times. During this second rifting phase (earliest part), a thick and cyclic succession of evaporitic deposits accumulated in the southern Rhine graben (COURTOT *et al.*, 1972; SITTLER & SCHULLER, 1988; BLANC-VALLERON, 1991), like in the Rhone graben (TRIAT & TRUC, 1974; DUMAS, 1986; CURIAL, 1987; MORETTO, 1987). Evaporitic circumstances of deposition continued to prevail through both these major segments of the European rift system until the Early Rupelian. In the Rhine graben these severely restricted, largely continental environmental conditions were suddenly and apparently simultaneously terminated on a graben-wide scale and replaced by an environmental complex of overall transgressively - regressively evolving marine facies during the later part of the Rupelian. In most of the Rhine graben marine marls and clays accumulated under humid climate conditions (DOEBL & TEICHMÜLLER, 1979). In succession, transgressive Foraminiferal Marls, anoxic Fish Shale and increasingly brackish regressive Meletta Beds were deposited in this rift segment during the Late Rupelian. In the Mainz basin and in the adjacent northernmost extension of the Rhine graben proper also significant sequences of fossiliferous "Meeressand" and "Schlicksand" accumulated. Deposition of all these strata occurred while rifting continued (ROTHAUSEN & SONNE, 1984; SISSINGH, 1998).

Within the Rhenish massif, the inception of the small Neuwied basin (near Koblenz) may have began later than

the initial development of the Rhine graben, i.e. during the (Early) Priabonian (MEYER, 1994; MEYER & STETS, 1996). S-N progressing rift propagation from the northern Rhine graben towards the post-Eocene Lower Rhine embayment (HAGER & PRÜFERT, 1988; ZAGWIJN, 1989; GELUK, 1990) may thus have initiated concomitantly diachronous, rifting-related subsidence and deposition in the north-western branch of the Rhenish Triple Junction. In the newly developed Neuwied basin, continental clays with lignite and conglomerates accumulated in Priabonian to Early Rupelian times, whereas in the Lower Rhine embayment marine clays and sands accumulated under influence of an overall NW-SE directed marine ingression during the Early Rupelian (HAGER & PRÜFERT, 1998). In the latter basin, similarly transgressive deposition continued during the Late Rupelian after a short-lived break in sedimentation. Contemporaneously in mid-Rupelian time, accumulation of brackish-water, lagoonal deposits was initiated in the region of the Neuwied basin (KADOLSKY, 1975; MEYER & STETS, 1996). However, saline communication does not seem to have existed with the Lower Rhine embayment. Most probably, marine ingression occurred from the Mainz basin area in the south and from the Paris and Luxembourg basins in the west. Continued rifting apparently led to a Late Rupelian marine connection between the North Sea basin and the Rhine graben via the Hessen depression. Since the marine Rhine graben was also communicating with the marine North Alpine Molasse basin (BERGER, 1996; SISSINGH, 1997, 1998), a long marine corridor was established between the North Sea and the peri-alpine basins during the later part of the Rupelian. Within this corridor, a southward and a northward current system occurred during respectively the early part and the middle to late part of the Late Rupelian (MARTINI, 1990). At the very end of the Rupelian, both the Rhine graben and the Hessen depression may have become relatively isolated from the North Sea basin, since open marine facies in these rift basins did not persist until the beginning of the Chattian (RITSKOWSKI & RÖSING, 1977; MEIBURG & KAEVER, 1986). In the Rhone graben clastics and carbonates accumulated during the Late Rupelian, together with some evaporites. Such continental to lagoonal deposits also accumulated in the coeval graben system of the Massif Central. The occasional Rupelian marine influences noticed in the Rhone graben and Massif Central basins are assumed to have had their origin in the West Alpine Molasse basin. Marine influxes detected in the Marseille basin, however, are supposed to stem from the palaeo-Mediterranean (Tethyan) realm and to indicate the occurrence of a N-S trending saline passageway in the region of the boundary between the Iberian and Corso-Sardinan blocks.

II.3.- Central Europe

II.3.1.- Polish Lowland basin

In (Late) Rupelian times, the Polish Lowland basin was invaded by the sea again, after an episodic regression during the latest Eocene. However, the transgression did not extend as far as in Lutetian - Bartonian times. Clayey and calcareous quartz - glauconite sands were deposited. The sandy sediments yielded planktonic foraminiferal faunas (including *Globigerina officinalis*, *G. turritilina* and *Rotaliatina bulimoides*) closely resembling those of the septarian clay facies of north-western Germany. The associations are devoid of warm-water elements (ODRZYWOLSKA-BIENKOWA *et al.*, 1978b). Their absence is in agreement with the boreal affinity of the molluscan faunas (JAKUBOWSKI & WOZNY, 1996). Towards the south and the east, the marine depositional domain merged into successively brackish and lacustrine (coal-bearing Czempin Formation) environments of deposition.

II.3.2.- Carpatho-Pannonian region

II.3.2.1. General features

The northward propagation of the Apulian (Adriatic) plate resulted in the closure of the Rheno-Danubian flysch trough during the late Middle and Late Eocene. The major Late Eocene tectonic movements (which also resulted in the separation of a discrete Paratethyan marine biogeographic province from the Tethyan in the Early Oligocene; NAGYMAROSY, 1994) were associated with a further outward shift of the external thrust front of the Magura accretional prism (Fig. 17.3). In the Late Rupelian, predominantly turbidite sedimentation took place in the internal parts of the flysch basins, adjacent to the emerging mountain chain, while accumulation of dark muds, "menilites", and calcareous oozes, reflecting in general dysoxic to anoxic conditions, prevailed in the more externally located parts. As elsewhere in the Paratethys, the environmental conditions in the basins of the Carpatho-Pannonian region were characterised by (at least episodically) reduced salinities and anaerobic conditions in response to increased isolation from the Mediterranean (RUSU, 1988; RÖGL, 1998). The reduced salinity and the anaerobic/dysaerobic conditions were primarily a consequence of Late Eocene tectonics, since the uplift of mountain chains (Southern Alps, Dinarides, Balkanides, Pontides, Anti-Caucasus, Kopet Dagh) and the closure or narrowing of marine corridors resulted in reduction of salinity and changes in circulation patterns in the Paratethys basins (NAGYMAROSY, 1991).

II.3.2.2.- Regional aspects

The events around the Eocene - Oligocene transition reflect an "acceleration" of the geodynamic evolution of the Carpatho-Pannonian domain. This was accompanied by important changes in the palaeogeographic configuration of the depositional areas of the Outer Carpathian flysches and with a shift of basin axes towards the North European platform. However, no break in sedimentation occurred across the Eocene - Oligocene transition. In contrast to the Eocene successions, which were deposited below the CCD, the Oligocene sequences accumulated above the CCD, as evidenced by the occurrence of calcareous plankton associations (KSIAZKIEWICZ *in* RAKUS *et al.*, 1990). In the Krosno - Menilite basins (Zdanice - Subsilesian, Silesian, Dukla and Skole units; see Fig. 17.3) black, bituminous clays and cherts were deposited (Menilite beds) in euxinic environments, along with (subordinate) deposition of psammites of turbiditic origin and slided bodies of strata (RAKUS *et al.*, 1990). In the Dukla and Silesian units the Menilite Formation reaches a thickness of 300 m (BIELY, 1996). Menilite-type deposits are rare in the Magura basin. Here, the Early Oligocene successions are characterised by

fine-grained rhythmites (calcareous flysch) of the Malcov beds, which developed in the internal zone of the Magura unit (Bystrica, Krynica and eastern part of the Raca subunits), as well as in the Dukla unit and in the marginal zone of the Central Carpathian Palaeogene basin. Menilite-type claystones are found in the lower part of the Malcov sequences (BIELY, 1996). In general, the Early Oligocene palaeocurrents of the Outer Carpathian flysch basins were directed towards the north. However, they were oriented towards the south-east in front of the Western Carpathians. The Early Oligocene (Kiscellian) sequences of the Central Carpathian Palaeogene basin are characterised by cyclically-bedded (sandstone) flysches (probably represented by the uppermost part of the Zuberec and Biely potok formations, GROSS *et al.*, 1984), which were derived from emerged parts of the Central and Internal Carpathian zones (carbonates and crystalline rocks; RAKUS *et al.*, 1990). The Biely potok Formation consists of several metres of sandstones, but also some claystone interbeds or polymict conglomerates occur. The formation reaches its maximum thickness (900 m) in the Levocske vrchy mountains. In the east, in the Maramures region, the Late Rupelian is represented by the Valea Carelor beds, which were deposited in anaerobic environments. The occurrence of large olistoliths points to tectonic instability in this part of the Central Carpathians.

In the Hungarian Palaeogene basin, depocentres shifted towards the north-east. The deposition of (upper) bathyal successions (Buda marls) persisted from the Late Eocene into the early part of the Early Oligocene. It was followed by the accumulation of the dark, predominantly laminated, bituminous, argillaceous siltstones of the Tard Formation (BALDI, 1986; NAGYMAROSY, 1997). The upper part of the formation is composed of the partly brackish «Pteropoda and *Cardium lipoldi* beds» (BALDI, 1986; CSASZAR, 1997; BALDI *in* BERCZI & JAMBOR, 1998), which were deposited in anoxic, upper bathyal environments. In the Szolnok flysch basin the sedimentation of turbidites in the Eocene gave way to the deposition of alternating shales and sandstones in the Early Oligocene; some beds reflect gravity-induced sliding (NAGYMAROSY & BALDI-BEKE, 1993; NAGYMAROSY, 1998).

In the south-eastern parts of the Central Paratethys, reduced salinities (9 - 16 promille) and dysoxic conditions were common in the deeper-water environments of the eastern Outer Carpathian flysch basin (and its south-eastern extension, the Getic basin) and in the Transcarpathian flysch basin and its adjacent zone in north-western Transylvania. The overall tectonic setting of the Eastern Carpathian domains was relatively stable in the Early Oligocene, as compared with the Eocene. Extension may have played a role at the beginning of the Rupelian (BADESCU, 1998), with a N-S orientation in the Transylvanian basin (HUISMANS *et al.*, 1997). The presence of foreland-derived Dobrogean-type greenschist clasts in the western parts of the Eastern Carpathian flysch basin suggests some, probably compression-related changes in the adjacent accretionary wedge. These changes were accompanied by a westward shift of the platform source areas (foreland bulge), due to increased loading effects, (BADESCU, 1998). Two types of sediment successions were deposited in the Carpathian trough (SAULEA *et al.*, 1971): foreland-derived sapropelitic muds of the Lower Dysodilic Formation with "Kliwa-type" arenites in the external parts and argillaceous flysch ("Shaly Horizon") supplied from the emerging Carpathians in the internal parts (SANDULESCU & MICU, 1989). The basin was characterised by steep western and more gently-inclined eastern slopes; water depths did not exceed 400 - 500 m. Arenitic deposits (Birtu Sandstone) are known from the Transcarpathian flysch zone; they were supplied from source areas in the west (SANDULESCU & MICU, 1989). Coarse, siliciclastic sediments deposited in littoral environments (Gruia Sandstone) have only been preserved on the Transylvanian shelf; they pass into the deeper-water bituminous shales of the Ileanda Formation (RUSU, 1977).

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

The main bathymetric features of the Early Oligocene basin systems of the Eastern Paratethys were fairly similar to those before (SCHERBA, 1993; KOPP & SCHERBA, 1998). The isolation of the Central European (Central Paratethys) and Black Sea - Caspian basins (associated with overall moderately warm, humid climate conditions and estuarine water circulation patterns) resulted in recurrent episodes of stagnancy of parts of the water column, and, consequently, in the accumulation of anoxic sediments (POPOV & STOLYAROV, 1996). Such sediments are common in the Oligocene and Lower Miocene and referred to as "Maykopian facies" in the Eastern Paratethys. In part, they correspond to the so-called menilites of Central Europe. The (near-) isolation of the Paratethys domains from the world oceans in the course of the Rupelian led to a (temporary) decrease of surface water salinity. All over the Paratethys this Early Solenovian development resulted in conditions characterised by brackish salinities and by the widespread occurrence of endemic benthic faunas (molluscan associations with *Urbnisia* and *Janschinella*; ostracods of the *Disopontocypris oligocaenica* Zone; POPOV *et al.*, 1985) and of semi-marine to brackish associations of dinoflagellates and fish taxa. In Late Solenovian time, the communication with the world oceans was partly restored and the salinity changed from brackish to semi-marine, while anoxic sedimentation continued in the deeper-water environments. In the marginal parts of the basin systems silty and sandy sediments with benthic faunas were deposited (e.g., southern Ukraine, Volga - Don area, Ustjurt, Kyzylkum, "*Corbula*" beds of Georgia; POPOV *et al.*, 1993b). Palynological/palaeobotanical evidence suggests climatic conditions with a warming maximum in the Late Solenovian, accompanied by increasing aridisation, especially in the eastern parts of the Eastern Paratethys (AKHMETIEV, *in* POPOV *et al.*, 1993b). Common endemic taxa and correlative facies changes witness of the presence of free water-mass exchange between the Eastern and Western / Central Paratethys during the Late Rupelian. The position of the marine corridor, however, cannot be ascertained. Possibly, it was located along the northern margin of the Balkanides; its sedimentary expression may be hidden below the present-day fold and thrustbelt.

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

After a terminal Priabonian regression, almost the entire realm of the East European platform became part of an emerged landmass. Two major depressions developed in the southern parts of the landmass: the Dniepr - Donetz and Precaspian depressions. In the Early Rupelian (and, probably, also during the earliest Chattian) a marine connection existed between the Eastern Paratethys and the North Sea basin via the Dniepr - Donetz basin, the Pripyat Strait and the Polish Lowland basin, but non-marine conditions were predominant in this area during the Late Rupelian (Solenovian). Clayey shelf sedimentation prevailed in the Precaspian embayment prior to the end of the Oligocene, but these deposits have been preserved only in the Lower Volga area and in small depressions originating from later salt tectonics. The Ural High and the Ukrainian area were major positive topographic features of the East European platform, but no significant amounts of coarse clastics were supplied from these areas. In fact, sandy deposits time-equivalent to the fine-grained Maykopian facies have a limited distribution only in parts of the southern Ukraine and in the lower Don, North Ergeni and north-eastern Precaspian areas.

Most of the domains of the Scythian plate were characterised by clayey shelf sedimentation during the Oligocene. Uplift of the southern margin of the Scythian plate and / or the development of an island chain (?Crimea, West Caucasian island) caused the separation of the shelf areas from the deep-water, central parts of the Eastern Paratethys. Sandy sediments were deposited in a narrow zone adjacent to the developing relief. To the north, deep-water, anoxic conditions existed in the Indol - Kuban and Terek - Mangyshlak depressions. Late Rupelian clays reach thicknesses up to more than 500 m in the Indol-Kuban depression, but only a few metres of sediment accumulated in the central part of the Terek - Mangyshlak depression, where sedimentation did not compensate for subsidence (STOLYAROV, 1991).

II.4.2.2.- Turanian plate and Kopet Dagh

Apart from the South Mangyshlak depression, the territories of the Turan plate represented shelf areas with sandy to silty sedimentation in the north-eastern Precaspian, North Pre-Aral and Kyzylkum embayments, whereas predominantly clayey sedimentation took place in the Ustjurt, South Aral and Fore-Kopet Dagh areas (SOLUN, 1975). The Eastern Kopet Dagh was transformed into a highland, acting as a source area of sandy (eroded Eocene) sediments. The Western Kopet Dagh belonged to the outer shelf where characteristic "Maykopian" fine-grained deposits accumulated throughout the Oligocene. Reddish clays and evaporites were deposited in the easternmost part of the Kyzylkum embayment.

II.4.2.3.- Black Sea depressions and Greater Caucasian - South Caspian basin

The deepest part of the central Eastern Paratethys was subdivided into depressions, including those of the Western and Eastern Black Sea. In these depressions thick successions (up to 4000 to 5000 m) of Oligocene to Early Miocene ("Maykopian") sediments were deposited, as inferred from seismic surveys (TUGOLESOV *et al.*, 1985). During the Early Oligocene, the vast Palaeogene Greater Caucasian basin was characterised by turbiditic / flysch sedimentation, as known from the southern slopes of the present-day Greater Caucasus (Sochi area, sections in northern Georgia and northern Azerbaijan). The South Caspian depression formed the continuation of the Greater Caucasian basin. It is assumed that the provenance areas of the clastic sediments were constituted by the Shatsky Ridge, the Dzirula Massif and the Lesser Caucasian - Elburz landmass. The emerged Pontides/Lesser Caucasus/Elburz/Kopet Dagh regions separated the Eastern Paratethys domains from the Tethyan realm during the Late Eocene and the Oligocene. Conglomerates, sandstones and siltstones with rich benthic associations accumulated along these emerged regions. They represent deposition in shallow shelf environments situated along the southern margin of the Eastern Paratethys. These sediments can be traced from the Adzharo - Trialetic zone and the Akhalcicke depression (south-western Georgia) towards the Talysh (southern Azerbaijan). Deeper-water shelf environments with clayey sedimentation prevailed farther to the north-east. Especially in the later part of the Solenovian, marine incursions occurred, as inferred from benthic faunas and phytoplankton associations, but the position of the marine corridor(s) cannot be ascertained.

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi - Oman domains of the Arabian Peninsula

The Late Rupelian was characterised by sea level lowstand and emergence of the Arabian platform (POWERS *et al.*, 1966; LE MÉTOUR *et al.*, 1995). Marine sediments only accumulated along the margins of the south-eastern part of Arabia and in some areas subject to localised tectonic subsidence. Carbonate platform environments were reinstalled in the domain of the Muthaymimah trough, where major basin subsidence persisted until the Late Eocene (Al Jaww Formation; CHERIF *et al.*, 1992; LE MÉTOUR *et al.*, 1992). The Abat trough collapsed, as evidenced by the accumulation of thick successions of gravity-flow deposits (Tahwah Formation; WYNS *et al.*, 1992; LE MÉTOUR *et al.*, 1995). In the Dhofar region, SSW-NNE oriented extensional tectonics related to the opening of the Gulf of Aden, resulted in the development of elongated basins, in which initially siliciclastic and later on carbonate sequences were deposited (Ashawq Formation; PLATEL & ROGER, 1989; ROGER *et al.*, 1992; PLATEL *et al.*, 1992d).

II.5.2.- Israel

Uplift of the Arabo - Nubian landmass had resulted in a pronounced NW-directed shift of the coastline relative to the Lutetian in Rupelian time. The concomitant regression caused the emergence of the entire Negev region and of parts of Judea and Galilee. Marine sedimentation was confined to a SW-NE trending basin, situated between an uplifted area off the present-day

Mediterranean coastline in the west and the Arabo-Nubian landmass in the east. In the west, i.e., in the depositional area of the present coastal plain region, marls of the Bet Guvrin Formation were deposited. Reefal limestones (Lakhish Formation) developed in the Shefela region, while sandy, dolomitic limestones (Fiq Formation) accumulated in the area east of the Sea of Galilee.

II.5.3.- Egypt

In the Early Oligocene, the northward progradation of the shoreline had advanced to about 30 degrees N (Brown & Kraus, 1988). Subsurface data evidence the occurrence of open marine depositional environments in the northern part of the Western Desert (shales with subordinate limestone interbeds of the Dabaa Formation). The sedimentary record of Dabaa well no. 1 shows 275 m of shales with sandstone and limestone intercalations, which belong to the *G. opima opima* Zone (Sheikh & Faris, 1985). Sequences which accumulated in the continental domains are represented by the deltaic Quatrani, the fluviatile Gebel Ahmar and the lacustrine Nakheil formations. Their deposition was associated with the development of river systems which drained uplifted areas in the Red Sea and Uweinat regions (Issawi *et al.*, 1999).

II.5.4.- Tunisia

The predominantly clastic sediments of the Rupelian of Tunisia allow to distinguish two major types of depositional sequences, based on their regional distribution and overall lithological characteristics. The first is typified by the sandy to clayey sequences of central, eastern and north-eastern Tunisia (lower part of the Fortuna Formation; Burollet, 1956); they are time-equivalent with the more eastward deposited Ketatna and Salamboo formations (Fournié, 1978). The second type of depositional sequences is composed of conglomerates, sands and clays. It is represented by the lower part of the allochthonous Numidian series (Rouvier, 1977; Yaich 1992, 1997). The Numidian deposits cover the northern part of Tunisia; more to the north they are found in the region of the "Galite Channel - northern Tunisian Plateau" (Tricart *et al.*, 1991). In central and north-eastern Tunisia the lithological composition of the Rupelian sequences (N1/N2 zonal interval; Ben Ismaïl-Lattrache, 1981; Ben Ismaïl-Lattrache & Bobier, 1984; Hooyberghs *et al.*, 1990; Yaich *et al.*, 1994; Yaich & Hooyberghs, 1996) strongly varies, both in space and time. Generally, relatively fine-grained, sandy to clayey or calcareous successions with gypsiferous interbeds accumulated. Towards the east, the amount of terrigenous clastics progressively diminished. Sedimentary and faunal characteristics reflect deposition in environments ranging from shallow-marine or lagoonal (south-east of the "North - South Axis") to deltaic (Yaich, 1997). The deltaic deposits were trapped in a NE-SW depression trending along the "Nebhana - Cap-Bon Axis". Bioclastic ramp deposits developed in the vicinity of the "North - South Axis". Locally, platform carbonates with nummulite associations accumulated. Deposition of open-marine successions was mainly confined to the Gulf of Gabes region (Yaich, 1997). Overall, the clastics were supplied to this region from westerly located hinterlands drained by W-E and, locally, NW-SE rivers systems passing into small deltas in the coastal areas. In front of the deltaic coastal zone beach barriers developed. The depositional domains of central Tunisia were bordered by emerged areas, such as a south-western landmass (comprising large parts of southern Tunisia), an uplifted area in the domain of the Gulf of Hammamet (extending as far as south of the south-westernmost part of Sicily), and a narrow ridge along the "Kef - Oued Zarga - Ghar el Melh Axis". The main provenance area of the Rupelian clastics deposited in the central and north-eastern Tunisian domains was the Tunisian Sahara, which was a continental realm from at least the Palaeocene onwards. The Rupelian of the Numidian basin consists of turbiditic successions comprising channel-fill sandstone, (subordinate) mudstone, channel-margin levee and overbank, interchannel and reworked levee deposits, as well as channel/lobe and pelagic/hemipelagic sediments with some turbidites and chaotic interbeds. The channels supplying the basin trended from the north to the south (Yaich, 1992).

In central and eastern Tunisia sedimentation was controlled by regional, SW-NE and NW-SE oriented extension (Haller, 1983; Turki, 1985; Yaich, 1986; Turki *et al.*, 1988). N 80 directions of synsedimentary extension prevailed along the "North-South Axis". The tectonic displacements resulted in a mosaic of mainly N 40 and N 140 oriented horsts and grabens (Turki, 1985; Yaich, 1986, 1997). The extensional stress regime was probably established in response to the opening of the Western Mediterranean basins (Dercourt *et al.*, 1986).

20.- EARLY BURDIGALIAN (20.5 - 19 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The mapped interval covers a time-window of about 1.5 Ma, from about 20.5 to 19.0 Ma (Fig. 17.1). It is characterised by calcareous nannoplankton associations indicative of the upper part of NN2 and the lowermost part of NN3. These nannoplankton associations allow to correlate the Early Burdigalian with the Eggenburgian and Sakaraulian regional stages of the Paratethys (see also ANDREYEVA-GRIGOROVICH & GRUZMAN, 1994; ANDREYEVA-GRIGOROVICH *et al.*, 1997). Micromammal associations correlative to the Early Burdigalian belong to zone MN3.

1.2.- Structural setting and kinematics

Major tectonic events occurring during the Late Oligocene - Early Miocene included the opening of the Gulf of Lions, Provençal - Ligurian, Valencia and Alboran basins in the Western Mediterranean domain. N-S directed rift propagation from the Rhone graben area resulted in separation of the Sardo-Corsican block from the main European plate from the "Middle" Oligocene onwards. Anti-clockwise rotation related to drifting of the block in a south-eastern direction commenced around the Oligocene - Miocene transition and seems to have lasted until the late Early Miocene to earliest Middle Miocene. These geodynamic events had some prominent coeval counterparts on the southern Peri-Tethys platform. In the Late Oligocene, the African - Arabian plate started to break up at the location of the present-day Gulf of Suez/Red Sea rift system. This "proto-rift" development was approximately coeval with the inception of sinistral displacements along the Dead Sea transform fault and succeeded by an Early Miocene synrift stage.

1.3.- Outlines of palaeogeography and palaeoenvironments

In the Late Oligocene (Chattian), open-marine conditions had become reinstalled in most domains straddling the Tethyan - Paratethyan transition zones after the Early Oligocene episode of often reduced salinities associated with severed connections with the world ocean systems. In Early Burdigalian time, there was a seaway connecting the basins along the collision zone from the Indo-Pacific to the Atlantic, as evidenced by facies and faunal distribution patterns. Various faunal elements from the Indo-Pacific, including large Pectinids (*Chlamys gigas),* migrated all along the Alpine foredeep (for a review, see RÖGL, 1998). Most probably, the connection between the Paratethys and the Indo-Pacific was located across eastern Turkey and Iran. However, the position of the marine corridor between the Eastern and Central Paratethys basins has not been ascertained. It is assumed that a narrow corridor may have existed south of the Moesian platform rather than a corridor across the platform near the Dobrogea massif ("Berlatsky corridor" of authors). The latter hypothesis lacks positive evidence because of the complete absence of Burdigalian marine deposits in the area. Shallow marine, relatively coarse terrigenous clastic sedimentation prevailed in the peri-Alpine molasse basins; farther to the east, in the Outer Carpathian "residual flysch trough", clays and turbidite sequences accumulated in deep-water basins. In some of the Central and Eastern Paratethys domains (for instance in the Carpathian and Caucasus - Kopet Dagh basins) dysoxic to anoxic conditions resulted in the deposition of laminated clayey successions ("upper menilites" and "upper Maykopian facies"). In some of the eastern areas sedimentary manganese ore deposits developed. In contrast to the open-marine conditions in the interconnected basins of the Atlantic - Indo-Pacific seaway, large-scale regression prevailed in the northern part of the Eastern Paratethys, resulting in the emergence of the Volga - Don area and Precaspian - Turan domain, excluding the Fore-Kopet Dagh Gulf. Also the sequences of the European rift system witness of a decreasing marine influence, reflecting the break-up of a marine connection between the Tethyan realm and the North Sea basin in Burdigalian time. Overall uplift of the western part of the northern platform equally caused the termination of marine conditions in the Paris basin at about the Aquitanian - Burdigalian transition.

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On the Iberian peninsula, intra-montane fluvio-lacustrine conditions persisted, accompanied by deposition of evaporites. In the south, large parts of the Iberian block had submerged since Late Rupelian time. The inception of rifting on the southern Peri-Tethys platform in the Late Oligocene was followed by the accumulation of open marine sediments in the silled Gulf of Suez - northern Red Sea basin and in the southern Red Sea - Gulf of Aden basin. Faunal evidence ascertains a marine connection between the Gulf of Suez / northern Red Sea basin and the Mediterranean realm across the Suez sill, while the southern part of the Red Sea domain was connected with the Indian Ocean. The northern and southern parts of the Red Sea domain were most probably separated by a still emerged isthmus in the central part of the evolving rift system.

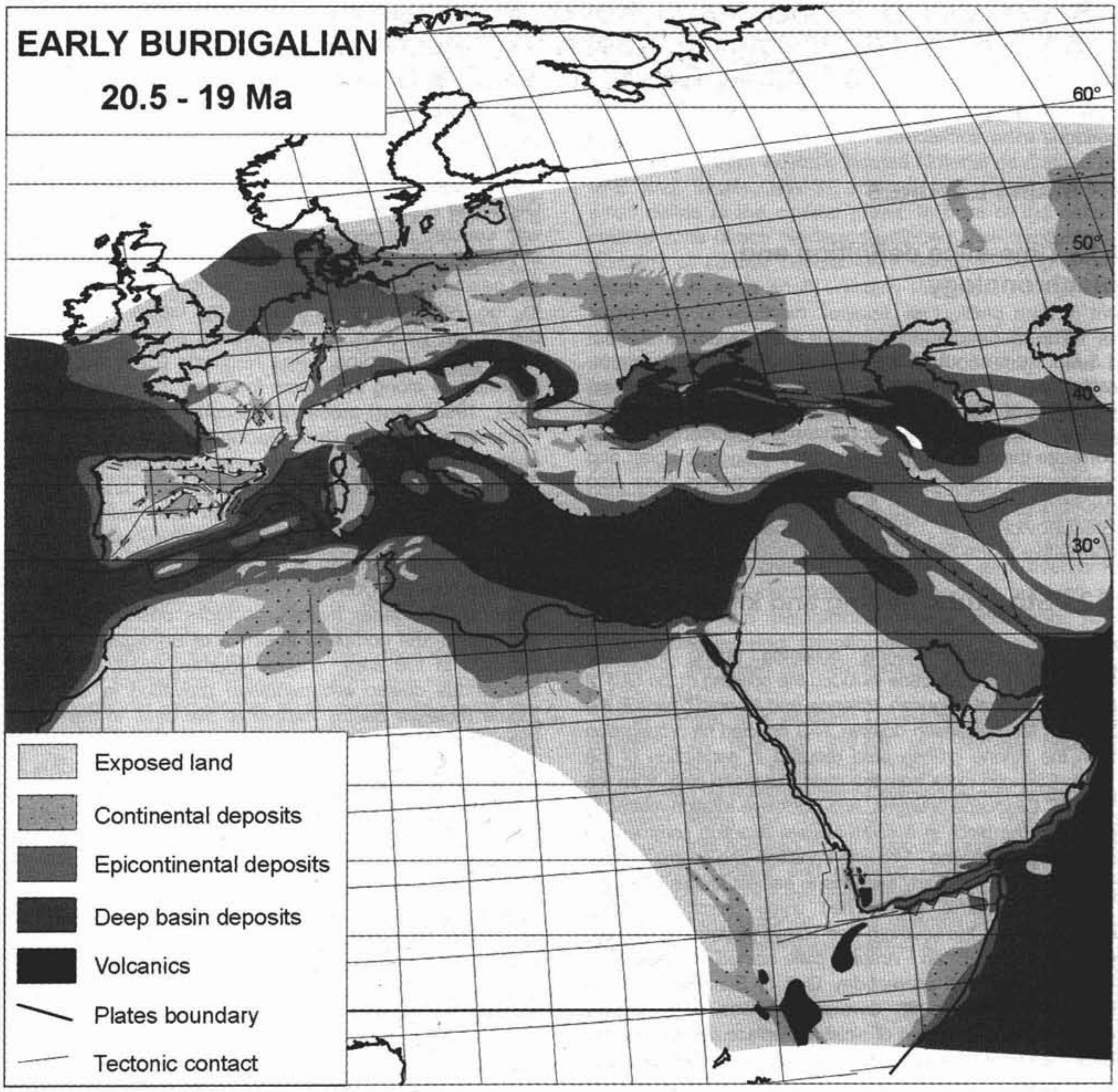

Fig. 20.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Burdigalian.

II.- DESCRIPTION OF DOMAINS

II.1.- South-western Europe

II.1.1.- Iberian Peninsula

The overall extension which had started in the Oligocene continued throughout the Neogene. However, the rifting of the Valencia trough ceased prior to the Middle Miocene (SANZ DE GALDEANO, 1996). Compressional deformation affected the southern margin of the Iberian plate, which resulted in the northward progradation of the Betic orogenic zones. The approximately N-S oriented stresses were also transmitted to the more northern regions, e.g., the Central Range system and the Pyrenees. This led not only to increased subsidence rates of the pre-existing interior, continental basins, but also to the origin of new, small basins, particularly so in the north-eastern parts of central Iberia. Sinistral displacements along the Plasencia fault persisted; reactivation of faults occurred (also) in the western, marginal areas of Iberia (ANTUNES *et al.*, 1999). The Early Burdigalian sedimentary record of the Ebro, Tagus, Duero and Calatayud - Daroca basins represents accumulation of alluvial deposits along the basin margins, interfingering with lacustrine sedimentation in the central parts of the basins. This general, lateral change in depositional facies is exemplified by the sequences of the Ebro, Calatayud and Tagus basins, where the lacustrine successions are mainly composed of evaporites (CALVO *et al.*, 1996; ORTI & SALVANY, 1997). In the Valles - Penedes basin in north-eastern Spain alluvial fan deposits accumulated in an environmental setting including minor lakes (CABRERA & CALVET, 1996). Along the western margin of Iberia (Portugal) continental and shallow-marine, littoral deposition occurred (ANTUNES *et al.*, 1999).

II.1.2.- Aquitaine basin

At the Oligocene - Miocene transition, an important phase of compressional tectonics induced (renewed) uplift of pre-existing anticlinal structures, extrusion of diapirs and major overthrusting of the northern Pyrenean orogenic front. These developments resulted in erosion of emerging reliefs and, subsequently, deposition of detrital successions along the southern margins of the Aquitaine basin during the Early Miocene. A major transgression occurred in the Aquitaine basin during the Early Burdigalian. Neritic, often bioclastic sediments were deposited. Some coral reefs developed in the northern and southern parts of the basin. They represent the last episode of reef growth in the Aquitaine basin. The rich and diversified (malaco)faunas of the Early Burdigalian (including the Burdigalian stratotype successions of Saucats and Léognan, and the "crags" of the Bordeaux area; POIGNANT & PUJOL, 1978; CAHUZAC *et al.*, 1997) indicate tropical climate conditions. Due to the westward progradation of shelf deposits deeper-water, pelagic sedimentation was confined to the western part of the basin. In the south-westernmost part of the basin mainly marly sediments accumulated in littoral environments. These sediments were deposited downstream in a W-E oriented canyon, which had been formed during the eustatic sea level lowering at the Rupelian - Chattian transition. Widespread lagoonal deposition of marls with *Crassostrea aginensis* occurred farther to the east. Continental successions are mainly composed of molasse-type sediments ("Molasse de l'Armagnac") with some local lacustrine interbeds (CROUZEL, 1956). In the eastern basinal area, the Burdigalian strata overlap widespread lacustrine limestones ("Calcaire gris de l'Agenais") of (latest) Aquitanian age.

II.2.- Western Europe

II.2.1.- Southern North Sea basin

Miocene deposition was episodically resumed in the southern North Sea basin in sandy and clayey facies after the Aquitanian phase of tectonic uplift. All over this southern part of the basin episodic accumulation of various glauconitic and shelly, shoreline-related sandstones and marine offshore clays predominated during Miocene and Pliocene times under influence of global changes of sea level and an occasional, Middle Miocene phase of tectonic uplifting of the bordering Ardennes and Brabant areas (LETSCH & SISSINGH, 1983; VINKEN, 1988; VANDENBERGHE *et al.*, 1998).

II.2.2.- Alpine foreland basins

At the beginning of the Burdigalian, which was a period characterised by a warm to subtropical climate north of the Alps, a widespread and rapid eustatic marine transgression established a wave-and tide-controlled seaway in the North Alpine Molasse basin (BERGER, 1996, SISSINGH, 1997). The marine ingression entered the basin both from the west and the east. The sea covered a considerable portion of the future Jura mountains and likely transgressed the Rhine graben from the south. Altogether, the Early Burdigalian peri-alpine seaway connected the Western Mediterranean basin with the Carpathian foredeep, presumably through the development of a narrow and short-lived corridor more or less in the region of the previous West Alpine Molasse basin. The inference of a strongly tide-influenced depositional realm with at least mesotidal regimes implies that the seaway was clearly connected with the world ocean. Water depths reached distally some 100 m (BERGER, 1985); however, within the basin tidal banks, shoals and islands occurred. Along the northern Alpine thrust wedge conglomeratic fan deltas continued to grow at the mouths of northward-discharging intra-Alpine streams.

After a brief, tectonics-related regression (deduced from a palaeosoil horizon), the Late Burdigalian interval was characterised by a rise in relative sea-level and by an enhancement of clayey deposition, encroaching depositional cycles of fan-deltas and other proximal deposits. Between the fluvial distributary systems, which are characterised by conglomeratic and sandy channel fills and sheet flows, tidally-influenced coastal facies including beaches, tidal sand waves and channels as well as intertidal sand flats dominated. These facies are bordered by a near-shore facies belt, which includes subtidal shoals with megaripples and intershoal swales and which is farther offshore replaced by glauconitic and pebbly coquina banks. At the northern margin of the North Alpine Molasse basin

sandy beaches and channelised tidal flats were established along rock coasts in the region of the present-day Jura (Homewood, 1986; Homewood & Allen, 1981; Homewood *et al.*, 1989). Farther to the east, the northern basin margin was bordered by shoreline cliffs which were cut into Late Jurassic limestones during the early part of the Late Burdigalian. In the eastern part of the North Alpine Molasse basin, a mainly west-directed transgression claimed first in earliest Burdigalian time the area closest to the Alpine nappes, and, thereafter, onlapped from there northwards an elevated region that existed throughout the Early Burdigalian. Thrust-loaded subsidence controlled a general, basin-wide shift of the depocentre to the north, till near the present-day basin axis (Sissingh, 1997).

Around the Eggenburgian - Ottnangian transition ("Middle Burdigalian"), the entire eastern part of the basin was inundated by a tide-influenced "epicontinental sea". Subsequent Burdigalian sedimentation occurred in marine to brackish and continental facies. Whereas during Chattian to Middle Burdigalian times the axial sediment transport was directed to the east, it reversed towards the end of the Burdigalian. Thus an easterly, south-westward-directed fluvial system which discharged in the west in a marine basin was ultimately established (Büchi & Schlanke, 1977; Unger, 1986; Doppler, 1989; Reichenbacher, 1993).

II.2.3.- European rift system

The Chattian of the Rhine graben is represented by a sedimentary series which successively includes the predominantly brackish to limnic *Cyrena* Beds, the Freshwater Beds and a substantial portion of the *Cerithium* Beds. The sequence is composed of calcareous sediments, which were deposited under variously saline environmental conditions (Rothausen, 1988). During the Middle Chattian accumulation of the Freshwater Beds, communication with the external marine realm was temporally terminated at both ends of the Rhine graben. Deposition of the overlying *Cerithium* Beds continued until the Aquitanian when these strata were succeeded by the *Corbicula* (or *Inflata*) Beds, which stratigraphically range up into the Middle Burdigalian. During the accumulation of the latter beds, the Rhine graben was a low-energy lagoon which experienced at times a poor water circulation with concomitant anoxic bottom conditions. In the vicinity of the Heidelberg subbasin, salt and anhydrite accumulated during the Early Burdigalian. The presence of brackish-water and more normal-marine depositional environments in the Rhine graben during the Chattian-Burdigalian interval necessitates the inference of marine passages between the Rhine graben and an external, marine realm. However, the identification of these corridors is as yet relatively conceptual. During the Early Chattian, the northward - directed current system in the saline passageway connecting the North Sea basin with the North Alpine Molasse basin, via the Hessen depression and the Rhine graben, continued more or less its existence. However, following in time the disappearance of the remnant marine environments in the North Alpine Molasse basin at the beginning of the Chattian, this communication system ceased to exist altogether in response to the Middle Chattian development of a barring high in the region of the Hessen depression (Meiburg & Kaever, 1986). It was presumably replaced by a saline corridor that intermittently connected the Mainz basin and the Lower Rhine embayment across the Rhenish massif during the Late Chattian to Aquitanian. This latter way of supplying the Rhine graben with saline water and fauna did not last necessarily until Early Burdigalian times. It seems that when marine conditions returned to the North Alpine Molasse basin at the beginning of the Burdigalian, a marine inlet developed between this basin and the Rhine graben. In co-occurrence with marine influences still continuing to arrive in the graben from the south, communication between the Lower Rhine embayment and the Rhine graben may have been re-established by Middle Burdigalian time. In the Neuwied basin (near Koblenz), Chattian deposition occurred, as far as known, in freshwater environments (Meyer & Sets, 1996). Only from the nearby Westerwald basin some remnant lagoonal deposits of Aquitanian age are known. In the southernmost part of the Lower Rhine embayment, lignite-bearing continental deposits were laid down in some unconformably succeeding sequences, ranging in age from Early Chattian to Burdigalian (Hager & Prüfert, 1988). In accordance with the foregoing conceptual interpretation concerning the existence of saline passages connecting the Rhine graben with external marine realms, it is assumed that marine ingressions have occurred in the overall swampy coastal lowlands of the southern Lower Rhine embayment in Late Chattian and Aquitanian time. However, correlative horizons of marine or brackish-water deposits are not known from the southernmost part of the basin.

In the Rhone graben, evaporitic deposits continued to accumulate during the Middle Oligocene. After an intra-Chattian break in deposition, the accumulation of sediment was resumed with sands containing *Miogypsina*. Upwards, these shallow marine deposits graded into brackish strata, which were soon replaced by lacustrine limestones and marls. The basal sandy beds indicate a short-lived incursion of the sea from the north. In general, the Chattian is assumed to have been mainly deposited in continental lagoons and lakes which were barred from the open-marine realm. The Late Chattian marine influence seems to have originated from the Rhine graben cq. North Sea basin (via the Lower Rhine embayment). At the very beginning of the Miocene, deposition was characterised by the accumulation of lacustrine limestones (Cavelier, 1984; Sissingh, 1998). Together with relatively marly deposits, they represent a palaeogeographic system including lakes. In the south, marine depositional circumstances occurred in the gradually northward prograding Gulf of Lions, an embayment which was bordered in the east by the south-eastward-rotating Sardinian - Corsica block. From the Aquitanian - Burdigalian transition onwards, the Rhone graben was again palaeogeographically strongly modified under influence of regional tectonics and a rapidly progressing eustatic transgression from the south. The southern part of the Rhone graben became part of a Lower Rhone archipelago (Demarcq, 1984). Concurrently, the Cucuron - Valensole basin developed east of the Durance fault. The northern part and most of the

central part of the Rhone graben were tectonically uplifted. Contemporaneously, uplift also occurred in the Massif Central. In this adjacent region, it initiated strong erosion and commensurate deposition of sands by an intra-graben system of S-N flowing rivers. As a consequence of this tectonic activity, the initial Burdigalian ingression of the sea from the Gulf of Lions never reached the Bresse graben and Valence basin. Within the structural confines of the Rhone graben, the marine incursion proceeded during the Early Burdigalian only till some distance south of Valence. Contemporaneously, however, a S-N directed transgression seems to have established a marine corridor between the Cucuron - Valensole basin and the North Alpine Molasse basin, closely along the western front of the Alpine orogenic wedge. During the Late Burdigalian, a seaway with tidal currents eventually developed from the Gulf of Lions across the Valence basin towards the North Alpine Molasse basin. *Avicenna* mangroves occurred along the newly defined shorelines of the Gulf of Lions (as they had been present since the (Late) Aquitanian). The North Alpine Molasse basin was also transgressed from the east; i.e. from the Central Paratethys domain.

II.3.- Central Europe

II.3.1.- Polish Lowland basin

After the Palaeogene, the basin was subject to emersion. In concomitance with its basin-wide, persistent continentalisation only continental, lacustrine, or brackish - lagoonal sediments were deposited in the Polish Lowland basin during the Neogene (PIWOCKI & ZIEMBINSKA-TWORZYDLO, 1997).

II.3.2.- Carpatho-Pannonian region

II.3.2.1.- General features

Geodynamic processes which influenced the Early Miocene evolution of the Eastern Alpine - Carpathian - Pannonian realm included the Alpine subduction and the north-vergent compression of the Adriatic promontory, which involved its collision with the North European platform. The collision of the Eastern Alpine orogenic domains with the Bohemian massif had initiated folding and thrusting in the Carpathian accretionary prism since the Late Eocene, as well as lateral extrusion of the lithospheric Alcapa fragment (Western Carpathian and Pelso megaunits) towards the north-east during the Oligocene - Early Miocene (CSONTOS *et al.*, 1992; RATSCHBACHER *et al.*, 1989, 1991a, b; TARI & HORVATH, 1995). In the Western Carpathians this resulted in the subduction of the Penninic-type units south and north of the Pieniny Klippen belt (see Fig. 17.3). In terms of palaeobiogeography and palaeoenvironmental evolution, the evolving Outer (Western, Eastern, and Southern) Carpathians, as well as the intra-Carpathian domains belonged to the Central Paratethys. In Early Burdigalian (Eggenburgian) time, marine connections existed between the latter region and both the Western and Eastern Paratethys, as evidenced by fauna and flora. In fact, the Eggenburgian was characterised by a major transgression which occurred with a concomitant immigration of marine biotas from the Atlanto-Mediterranean and Indo-Pacific bioprovinces (e.g., BALDI, 1979; RÖGL & STEININGER, 1983; RUSU, 1988; RÖGL, 1998). Characteristic littoral marine sediments with large Pectinids (*Chlamys gigas*) were deposited over large parts of the Central and Western Paratethys, ranging from the Transylvanian basin as far as the eastern North Alpine Molasse basin. The faunas witness of warm and fairly uniform climate conditions during the Early Burdigalian. The intra-Carpathian and Outer Carpathian basin systems were interconnected via the East Slovakian basin. In the former systems (Transylvanian basin and part of the Pannonian region) predominantly marine clays, marls and sandstones accumulated. Flysch deposition played an important role in the Outer Carpathian basin systems, now incorporated in the Outer Carpathian nappe piles.

II.3.2.2.- Regional aspects

a.- Outer Carpathians

The Early Miocene, Eggenburgian collision induced thrusting of the Northern Calcareous Alps and Rheno-Danubian Flysch over the internal zone of the Alpine foredeep and the Waschberg zone (see Fig. 17.3). Palaeostress reconstructions point to N-S compression (PERESSON & DECKER, 1997). The Early Miocene (Eggenburgian and Ottnangian) autochthonous sediment successions in the Alpine - Carpathian junction area are composed of sandstones, siltstones and clays and reach a maximal thickness of about 2500 metres (JIRICEK & SEIFERT, 1990). The earliest Miocene thrust front of the Outer Carpathian accretionary wedge was only active along the south-eastern margin of the Bohemian massif (Magura and Dukla units of the Northern Carpathians; OSCZYPKO, 1997). After a late Early Miocene (intra-Burdigalian, i.e. Ottnangian) compressional tectonic event, the Outer Carpathian active thrust front shifted to the front of the Subsilesian and Silesian units in the north, and to the inner Moldavides (frontal parts of the "Convolute Flysch" and of the Macla and Audia nappes) in the eastern and south-eastern parts of the evolving Carpathians (SANDULESCU, 1988; MICU, 1990; KOVAC *et al.*, 1995).

Deep(er)-water, high-energy, turbiditic sedimentation was characteristic for the "residual" flysch troughs of the Outer Carpathian Silesian - Subsilesian and Skole - Skiba - Tarcau zones of the Northern and Eastern Outer Carpathians (Middle/Upper Krosno, Polyanytsa, Vinetisu and Podu Morii formations). The turbidite sequences reach thicknesses varying from 500 to 1500 m (KOSZARSKI *et al.*, 1995; ANDREYEVA-GRIGOROVICH *et al.*, 1997; OSZCZYPKO, 1997). Coarse-clastic sediments (gravity flows) were derived from either external platform areas (e.g., Gura Soimului Formation), or from already uplifted internal parts of the Carpathians, as follows from the occurrence of cannibalised older flysches (e.g., Slon Member; see also SLACZKA & OSZCZYPKO, 1987). Flexural bending below the internal parts of the Eastern Carpathian foredeep resulted in the deposition of coarse to fine-grained clastics in shallow-water and high-energy environments and of evaporites in lagoonal environments. In the north-eastern parts of the Outer Carpathian basin systems the lower part of the Early Miocene successions, now incorporated in the Sambor - Rozniatov nappe (ANDREYEVA-GRIGOROVICH *et al.*, 1997), may reach

a maximum thickness of 3000 m (Sloboda/Dobrotiv conglomerates and Stebnik Formation). The sediments of the Subcarpathian folded Neogene (Gura Soimului and Salt formations; Pietricia - Plesu member) are considered to have been deposited in the southward extension of the shallow-water to lagoonal depression of the internal parts of the Eastern Carpathian foredeep. The composition of the Early Miocene conglomerates indicates provenance from the uplifted margin of the East European platform (SANDULESCU *et al.*, 1981; MARUNTEANU, 1985).

The Early Burdigalian differentiation of the various depositional environments in the southern parts of the Eastern to Southern Carpathians is reflected by (from more internal to more external domains): a) flysch sequences of the upper part of the Fusaru - Pucioasa unit of the Tarcau nappe, overlain by finely-bedded argillaceous shales and menilites (bituminous cherts), b) sediments of the upper part of the "Bituminous unit" (with a limited distribution in the Tarcau domain, but widely-distributed in the area of the Marginal folds unit and the Subcarpathian nappe) composed of flysch-type deposits, sandstones ("Kliwa sands"), menilites, micro-conglomerates and coarse-grained sandstones with marl interbeds, and c) arenites (deposited in the northern part of the South-Carpathian foredeep). In these internal to external, deep to shallow-marine environments, the Early Burdigalian sedimentation ended with the accumulation of evaporites (gypsum and salt).

b.- Intra-Carpathian region

The Early Miocene "Alcapa drift" was enhanced by the Alpine collision and the Carpathian subduction pull. The transpressional regime of the Eggenburgian was characterised by N-S compression in the Eastern Alps (PERESSON & DECKER, 1997). Towards the east, i.e., in the Western Carpathians, the palaeostress direction changed. Here, the NW-SE oriented axis of main compression mirrors the Miocene counterclockwise rotation of the Alcapa microplate over 60 - 80° (MARTON *et al.*, 1995; KOVAC & TUNYI, 1995; MARTON & FODOR, 1995; KOVAC & MARTON, 1998). The compression was associated with mainly ENE-WSW right-lateral displacements and SW-NE directed thrusting. In the northern marginal areas of the central Western Carpathians wrench faulting led to the development of elongate compressional basins, which were filled with 200 to 500 m of shallow-marine deposits (NEMCOK *et al.*, 1989; KOVAC *et al.*, 1994, 1995, 1997; FODOR, 1995). In the Transdanubian Central Range zone (TCR), NW - SE compression can be inferred from right-lateral strike-slip displacements and SW - NE trending thrusts (CSONTOS *et al.*, 1991; FODOR *et al.*, 1992). The TCR zone was uplifted in response to compression; sedimentation continued to the east, in the retroarc basin on top of the Bükk zone (TARI *et al.*, 1993). N-S to NNE-SSW directed compression (KOVAC *et al.*, 1995) occurred in the eastern parts of the Alcapa domains (north-western part of the Transcarpathian depression). This led to the gradual disintegration of the Oligocene - Early Miocene forearc basins in association with thrust deformations along the margin of the overriding plate (PLASIENKA *et al.*, 1998). In the course of the Early Miocene, i.e., during the Ottnangian, presumed dextral shear between the Pienyni Klippen Belt and the central Western Carpathians enabled the counterclockwise rotation of the Alcapa microplate (KOVAC & MARTON, 1998). The same dextral shear motions induced the opening of the East Slovakian basin in latest Early Miocene (Karpatian) time.

In the areas belonging to the Tisza - Dacia microplate (Fig. 17.3) the earliest Miocene thrusting and right-lateral strike-slip displacements probably had similar orientations as those in the Alcapa domains, as inferred from the pre-Tertiary units of the Mecsek mountains (CSONTOS *et al.*, 1991; CSONTOS & HORVATH, 1995). However, in the northern margins of the Transylvanian basin and in the Apuseni mountains a change of palaeostress orientation from NW-SE to NNE-SSW occurred at about the Oligocene - Miocene transition. This change may have post-dated the clockwise rotation of the Tisza - Dacia microplate. In summary, the available data indicate that palaeostress orientations were N-S in the Eastern Alps and NW-SE in the Central Western Carpathians, the Transdanubian Central Range and in the Bükk and Mescek regions. N-S to NE-SW trending compressional tectonics (HUISMANS, 1997; MATENCO, 1997) initiated north to north-east displacement of the Tisza - Dacia microplate. At its front, the "Convolute Flysch" and the Macla and Audia nappe systems were obliquely thrusted in response to intra-Burdigalian tectonics associated with transtensional - extensional deformations in the southern Carpathians (SANDULESCU, 1988).

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

In the Eastern Paratethys the first biogeographic indications for Mediterranean influence can be inferred from the lowermost Miocene successions. The appearance of warm-water Indo-Pacific faunal elements (foraminifera, molluscs, the fish genus *Alepes*) suggests the existence of a marine corridor through eastern Turkey (ERUNAL-ERENTOZ, 1958; LÜTTIG & STEFFENS, 1976) and Iran during the early part of the Early Miocene. After a widespread, short-term transgression at the beginning of the Miocene, regressive conditions became predominant in the domains of the Eastern Paratethys in the course of the successive part of the Early Miocene. This is especially evident in the northern parts of the Eastern Paratethys, i.e. in the southern Ukrainian area and in the Volga - Don, Ustjurt and Aral regions. As in the Oligocene, most of the successive Early Miocene deep-water environments were characterised by clayey sedimentation under anoxic conditions («Maykopian facies»), associated with an estuarine circulation system. Anoxic clays rich in organic matter and pyrite accumulated during the Early Miocene (including the Sakaraulian; POPOV & STOLYAROV, 1996). These sediments cannot be dated with certainty because of the absence of age-diagnostic fossils. The supposedly time-equivalent, rich and diversified shallow-water molluscan associations from the southern part of the Transcaucasus area reflect a climatic optimum during the middle Early Miocene. These associations include Indo-Pacific immigrants, presently inhabiting tropical seas (*Fragum, Plagiocardium*). Salinities were close to normal. Palynological

evidence from the Sakaraulian stratotype area indicates an increase in subtropical and exotic elements (L.A. PANOVA, pers. comm.). Arid floral elements (e.g., *Ephedra*) from deposits which accumulated at the eastern margin of the Eastern Paratethyan realm indicate seasonal climatic conditions including dry summers (AKHMETIEV *in* POPOV *et al.*, 1993b). Both palynological and sedimentological data corroborate the assumed overall regressive nature of the Sakaraulian time-slice, as characterised by depositional shallowing and (increased) terrestrial input.

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

All domains of the East European platform, including the Precaspian area, emerged in response to the overall regression which started shortly after the beginning of the Miocene. Major positive topographic features, which acted as provenance areas for clastic supply, included the Ural Mountains, the Pre-Ural Highland, and the Ukrainian and Donetz lands. The Dniepr - Donetz and Pripyat depressions were transformed into alluvial lowlands with lakes.

The domains of the Scythian plate continued to represent the northern shelf areas of the Eastern Paratethys. Their northern, shallow-water environments were subject to the deposition of sands and aleuritic material. The Terek - Mangyshlak depression was filled up by sediments derived from the East European platform during the Late Oligocene. It was thus transformed from a deep-marine shelf basin with condensed sedimentation into a shallower marine basin. The southern part of the Scythian plate was occupied by outer shelf depositional environments. Deeper-water conditions comprising deposition of anoxic clays prevailed in the northern Crimea area and in the Fore-Caucasian basin. The Indol-Kuban depression corresponded to the deepest part of the northern shelf areas of the Eastern Paratethys in Sakaraulian time. In this basin "Upper Maykopian" (i.e. undivided Early Miocene) sequences reach thicknesses of up to 1000 to 1500 m. In the northern shelf areas overall sedimentary changes occurred from accumulation of muds towards deposition of sands and silts during the Sakaraulian. Concurrently, the oxygenation of bottom water improved and, consequently, benthic communities diversified and inhabited newly-developed biotopes.

II.4.2.2.- Turanian plate and Kopet Dagh

Shallow seas occupied the northern (Ustjurtian and Pre-Aralian) parts of the Turan plate during the early and the later part (Kotsakhurian or "Rzehakia time") of the Early Miocene. In between, i.e. during the Sakaraulian, continental environments prevailed in these areas. In contrast, marine, sandy to clayey Sakaraulian deposits accumulated in the South Mangyshlak, Karakum and Fore-Kopet Dagh domains. In the South Mangyshlak basin a change from shallow-water into deeper-water shelf environments occurred (POPOV & STOLYAROV, 1996). The Karakum - Fore-Kopet Dagh Gulf was separated from the main basin systems of the Eastern Paratethys by the large Tuarkyr Island. The latter emerged area had considerably increased in size since the Oligocene (Late Rupelian), in response to the regional regression of the Early Miocene. Farther to the east, shallow-water clastics accumulated in the Fore-Kopet Dagh Gulf, in which basin the middle part of the Aktepe Sands was dated as Early Burdigalian. These sands yielded warm-water molluscan associations, which are similar to those of the Sakaraulian (VORONINA *et al.*, 1993). The source area of the sands was the Kopet Dagh Land. Only the north-western part of the Kopet Dagh area was submerged. It corresponded to a relatively deep-marine shelf depression in which anoxic clays were deposited. Altogether, the lowlands and continental depressions of the southern West Siberia, Turgaj, South Kazakhstan, Tien Shan (east of mapped area) and Tadjic (east of mapped area) regions constituted a palaeogeographic complex comprising large fresh-water lakes ("Great Lake Time"), in which especially clayey, sometimes reddish sediments accumulated, as during the Oligocene.

II.4.2.3.- Black Sea depressions and Greater Caucasian - South Caspian basin

The Early Miocene bathymetric features of the central area of the Eastern Paratethys were inherited from the Oligocene (KOPP & SCHERBA, 1985). The deepest basinal parts correspond to the Western Black Sea, the Eastern Black Sea and the Greater Caucasian - South Caspian depressions. The deepest-water "Upper Maykopian" sediments accumulated in the eastern Azerbaijan part of the Greater Caucasian - South Caspian basin. Coeval uplift of the Central Caucasus resulted in an increased supply of clastic material into the Fore-Caucasian basin (Laba Sands, siltstones of the Olginskaja Suite). Outer shelf environments existed along the Black Sea coast (Tuapse area) and in the Kura basin. Here, silty to clayey sediments accumulated. These deposits were derived from source areas presently incorporated in the Lesser Caucasus ridge and in the Adzharo - Trialet ridge. A narrow and elongate, sandy shelf zone extended from Western Georgia (coastal area of the Black Sea) to the Middle Kura depression and the Talysh in Azerbaijan, along the northern margins of the Adzharo - Trialet High and the Lesser Caucasus. In this region sandy and gravely sediments with rich and diversified benthic associations (more than 100 species of bivalves recovered in central Georgia; POPOV *et al.*, 1993b) were deposited. As a result of continued uplift of the Lesser Caucasus, Talysh and Elburz chains (since the Oligocene), sandy sedimentation prevailed on the Transcaucasus southern shelf area.

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi-Oman domains of the Arabian peninsula

Intensified rifting in the Gulf of Aden led to the separation of the Arabian and African plates at the beginning of the Late Oligocene. This main phase of extensional tectonics was associated with the collapse of the Dhofar margin along major listric faults. The early Late Oligocene major event initiated the accumulation of

thick successions of slope deposits, which continued in the southern coastal area of Dhofar until Burdigalian times (calciturbidites, debris flows and olistoliths of the Mughsayl Formation). At the same time, the inception of high sedimentation rates of calcareous debris-flow deposits in prograding sedimentary carbonate prisms (Shuwayr Formation) occurred in coastal areas, situated along the Indian Ocean (PLATEL & ROGER, 1989; PLATEL *et al.*, 1992a, b, c; ROGER *et al.*, 1992). The latter deposits were derived from the south-eastern Oman flexured margin, which was located along the edge of the emerged Huqf - Dhofar palaeohigh. In the north-eastern part of the Arabian plate mixed slope deposits (conglomerates and megabreccias of the Tahwah Formation) accumulated all along the Batinah coastal plain region (WYNS *et al.*, 1992).

II.5.2.- Israel

The N-S oriented Dead Sea Transform was active since latest Oligocene time, which activity was closely associated with the opening of the Red Sea Rift system. However, the sinistral displacement along the Dead Sea Transform did not yet affect the transport of large amounts of clastics from the east. These clastics testify for the existence of huge fluviatile systems, which drained the Arabian craton via the Beersheba Canyon (clastics of the Hazeva Formation) and the Yezreel Valley (clastics of the Hordos Formation) towards the west. Contemporaneously, accumulation of marine marls of the Bet Guvrin Formation persisted in the Levantine domain, in the area of the present-day coastal plain of Israel.

II.5.3.- Egypt

A major regression occurred after the Late Rupelian. It was roughly contemporaneous with the updoming of the Arabo - Nubian massif and the inception of rifting in the Gulf of Suez / northern Red Sea region. From the Early Miocene onward, four tectonically-controlled depositional areas developed: 1) the northern Nile delta embayment, 2) the northern Western Desert, 3) the Cairo - Suez district and 4) the Gulf of Suez - northern Red Sea region (SAID, 1981, 1990). Aquitanian sediments attained only a limited distribution (SADEK, 1968) as compared with those of the Burdigalian. The Burdigalian deposits extend over large parts of northern Egypt and are also equally well-represented in the newly-formed Gulf of Suez, which was connected with the Mediterranean domains via the Suez sill (SALEM, 1976; SAID, 1990). The regional distribution of the marine Burdigalian sediments portrays the most widespread transgression of the Miocene in Egypt (SAID, 1990). Fluviatile (Gebel El-Khasab red beds) and deltaic (Moghra Formation) sequences accumulated in the northern Western Desert. These sequences are time-equivalent to shallow-marine limestones (Mamura Formation) deposited farther to the north-west. In wells drilled in the Nile Delta embayment up to about 1800 m of marine clays deposited in bathyal environments have been encountered (HARMS & WRAY, 1990). Thin successions of shallow-marine limestones and shales accumulated in the Cairo - Suez area. In the Gulf of Suez and the northern Red Sea areas thick successions of shales and marls (Lower Rudeis Formation) accumulated in the deep-water (bathyal) environments of the evolving rift system (e.g., QUDA & MASOUD, 1993). Concurrently, alluvial fans (Ranga Formation) developed along the margins of the northern Red Sea rift.

II.5.4.- Tunisia

The stratigraphic delimitation and correlation of the Burdigalian deposits of Tunisia is uncertain because of the lack of sufficient biostratigraphic control. In central Tunisia, the "Early Burdigalian" comprises the upper part of the fluviatile sands of the Fortuna Formation (BUROLLET, 1956), as well as the red deposits of the Messiouta Formation (SCHOLLER, 1933; BUROLLET, 1956). As such, the "Early Burdigalian" represents successions sandwiched in between the well-dated, Aquitanian sediments of the Fortuna Formation (HOOYBERGHS, 1992) and the Late Burdigalian sands and clays of the Grijima Formation (YAICH *et al.*, 1987; YAICH, 1988, 1991, 1997). It is correlative to the shallow marine platform carbonates of the Ketatna Formation of eastern Tunisia (FOURNIÉ, 1978), which, in turn, pass laterally into more marly sediments in the region of the Gulf of Gabes. In the area from central Tunisia to the east and north-east two main depositional domains adjacent to the uplifted areas in the south and south-west can be distinguished, based on the correlation of outcropping sequences and subsurface information. One included coastal and shallow-marine depositional environments in the region of the Gulf of Hammamet. The other comprised marine carbonate platforms in the area of the Gulf of Gabes. They developed under influence of an extensional stress regime which resulted in the development of W-E and NW-SE trending horsts and grabens (HALLER, 1983; YAICH, 1984; BLONDEL *et al.*, 1988; YAICH, 1997). Prior to the deposition of the Messiouta Formation, large parts of central Tunisia emerged (e.g., Kasserine Island, Gulf of Hammamet). As a consequence, W-E or SW-NE oriented river systems developed. They crossed vast alluvial plains, onto which red soils developed under tropical, humid climate conditions. Along the margins of pre-existing or newly-formed highs the soils were eroded. The erosion products were redeposited in adjacent lacustrine to lagoonal depressions (BLONDEL *et al.*, 1988), along with clasts derived from older formations, which resulted in the deposition of sands, ferrugenous red beds ("limons rouges"), gypsiferous clays and limestones.

21.- EARLY LANGHIAN (16.4 - 15.5 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The base of the Langhian corresponds to the first appearance of *Praeorbulina,* dated at about 16.4 Ma; the upper limit of the Early Langhian is ill-defined. It probably falls within nannoplankton zone NN5 (between about 15.8 and 15.5 Ma; Fig. 17.1). Nannoplankton associations indicate that the Early Langhian covers the upper part of zone NN4 and the lower part of zone NN5 (MARTINI, 1968; FORNACIARI *et al.*, 1996, 1997). NN4 (upper part) assemblages have also been recovered from the Late Karpatian, while the base of the Early Badenian corresponds to the base of zone NN5 (MÜLLER, 1974; LEHOTAYOVA & MOLCIKOVA, 1975; MARTINI & MÜLLER, 1975; NAGYMAROSY, 1985; MARUNTEANU, 1992; ANDREYEVA-GRIGOROVICH *et al.*, 1997; ANDREYEVA-GRIGOROVICH & SAVYTSKA, in press)). Consequently, the mapped interval in the Central Paratethys comprises the Late(est) Karpatian and Early Badenian in terms of regional stages. However, there is no consensus on the correlation between the Langhian and regional stages of the Eastern Paratethys. Rare nannoplankton floras from the Tarkhanian stage contain associations indicative of zone NN4 or NN5 (for a discussion, see RÖGL, 1998). Also the position of the next-higher, Tschokrakian stage relative to the Langhian cannot well be ascertained because of the lack of open ocean index species. Based on the combined results of biostratigraphy, magnetostratigraphy and radiometric dating of the uppermost Early Miocene and lowermost Middle Miocene it is proposed here (I.A. GONTSHAROVA) to consider the Late Tarkhanian to Early Tschokrakian interval as the equivalent of the Early Langhian for the Eastern Paratethys domains (see also STUDENCKA *et al.*, 1998). However, the Late Tarkhanian and Early Tschokrakian deposits differ in lithology and distribution. The selected interval mapped in the Eastern Paratethys corresponds to the Early Tschokrakian. All over the Peri-Tethyan realms terrestrial sequences correlative to the Early Langhian comprise micromammal associations indicative of zone MN5 or, incidentally, MN6.

I.2.- Structural setting and kinematics

The Mediterranean realm and the Carpathian - Pannonian region underwent one of the most important palaeogeographic / palaeoenvironmental reorganisations during the Neogene in response to intra-Burdigalian tectonics. Subsequently, the major north-vergent thrusting of the Eastern Alps ended, concomitantly with the inception of rapid, arc-parallel, eastward migration of foreland depocentres in the Outer Carpathian domain and the opening of the Pannonian back-arc basin system in latest Burdigalian to earliest Langhian time. In the Western Mediterranean the Early Miocene chain of events resulted in the counter-clockwise rotation of about 45 degrees of the Sardo-Corsican block until about the Early - Middle Miocene transition, and in oceanisation of the Western Mediterranean basin from the Early Miocene onwards. The Aegean area witnessed large-scale, south-westward-directed thrusting, coupled with the opening of extensional basins in the rear of the prograding fold and thrust belt of the Hellenides (not mapped). In the eastern domains of the northern Peri-Tethys platform intra-Burdigalian tectonic instability resulted in a basic reorganisation of basin configurations prior to the beginning of the Middle Miocene. Major events governing the development of the southern Peri-Tethys platform include the "mid-clysmic event" affecting the evolving Gulf of Suez rift basin around the Burdigalian - Langhian transition.

I.3.- Outlines of palaeogeography and palaeoenvironments

Fluvio-lacustrine and evaporitic sequences continued to be deposited during the Langhian in the intramontane basins on Iberia, while marine conditions

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persisted around the Iberian massif. In the North Alpine Molasse basin (Western Paratethys) marine sedimentation had come to a close near the Burdigalian - Langhian transition. It was replaced by accumulation of terrigenous, non-marine clastics predominantly derived from the emerging Alpine chain. Further to the east, deep-marine sequences were deposited in foreland basins along the emerging Outer Carpathians. These basins were connected with the contemporaneously developing intra-Carpathian back-arc basin system, in which locally deep-marine, partly dysoxic conditions existed. The Early Langhian basins of the Carpathian - Pannonian arc and back-arc system communicated with the Mediterranean realm through the "Trans-Tethyan Trench corridor". Connections between these basins and those of the Eastern Paratethys had ceased to exist. The Eastern Paratethys witnessed a regional transgression, particularly so in the north-eastern and eastern domains, as expressed by the contours of the North Caspian and Ustjurtian gulfs, although marine connections with the world oceans (via the Central Araksian Strait and a passageway across Iran/eastern Turkey) became restricted. This restriction resulted in the development of endemic faunas in the course of the Tschokrakian. Sequences with evaporitic intercalations accumulated east of the Caspian domain. The palaeogeographic

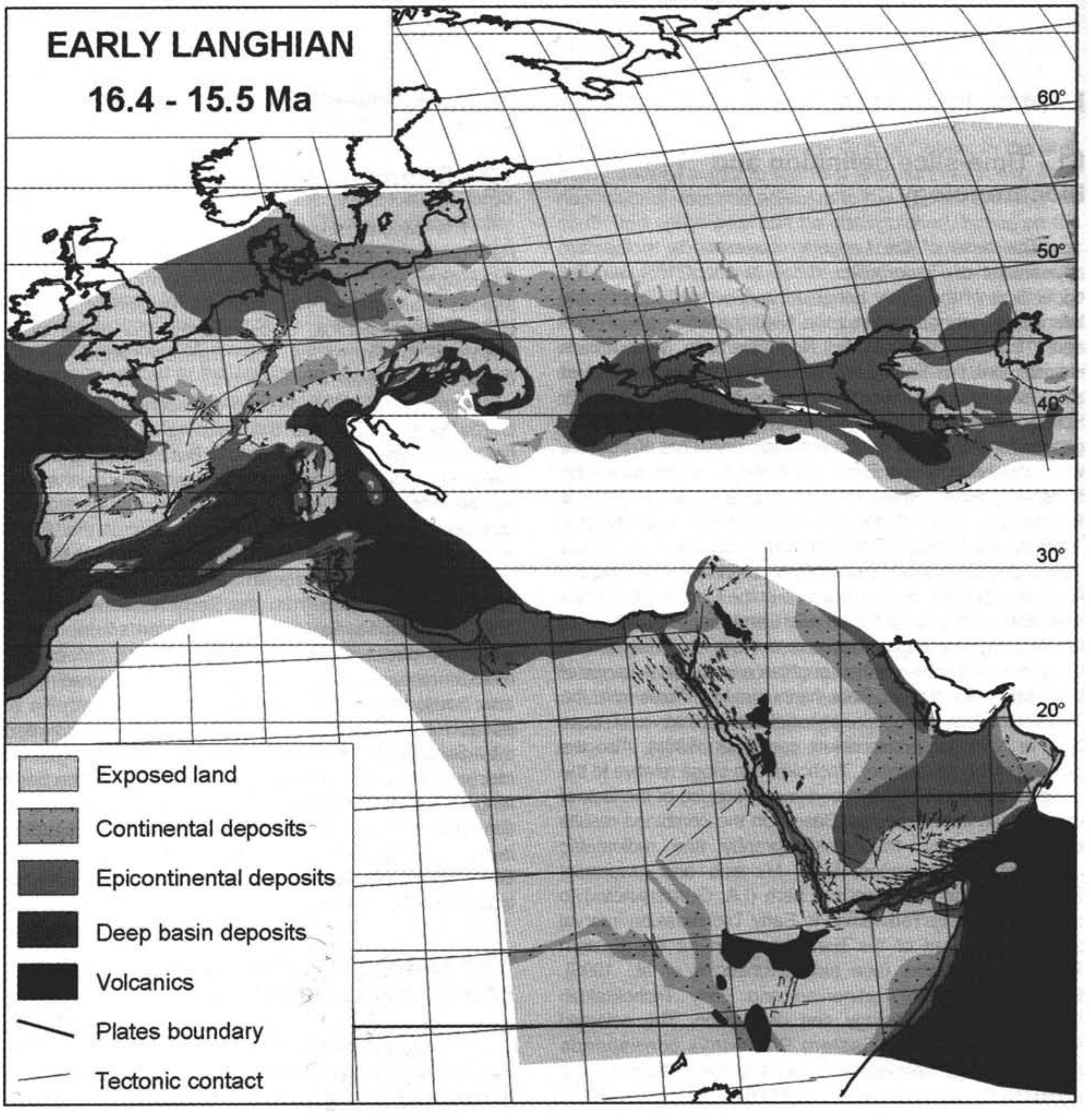

Fig. 21.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Early Langhian.

configurations and environmental conditions in the north (i.e. in the interconnected North Sea basin and the Polish Lowland and Dniepr - Donetz depressions) were roughly similar to those prevailing in the Early Burdigalian.

The Arabian platform was subjected to a major transgression in latest Burdigalian to Early Langhian times. In the northern parts of the Gulf of Suez - Red Sea rift system open marine, partly dysoxic to anoxic conditions prevailed in a configuration of silled subbasins; in the marginal areas carbonates or evaporites were deposited. Faunal evidence indicates a marine connection with the Mediterranean across the Suez sill. The Gulf of Suez - northern Red Sea basin was probably still separated by an emerged isthmus from the basins in the southern Red Sea area and the Gulf of Aden, which segment of the rift system was connected with the Indian Ocean. West of the Suez sill, as far as Libya, terrigenous - clastic sedimentation gave way to the deposition of shallow-water platform carbonates.

II.- DESCRIPTION OF DOMAINS

II.1.- South-western Europe

II.1.1.- Iberian Peninsula

During the Early Langhian, the major, westward displacements of the internal zones of the Betic Range came to an end. The associated compressional stresses, along with those persisting in the northern parts of Iberia (Cantabrian and Pyrenean Ranges) induced further uplift of the northern margin of the Ebro basin, as well as reactivation of pre-existing fault systems (Central Range system, Plasencia fault). Reactivation of faults resulted in increased accumulation rates of alluvial sediments in troughs fringing the margins of the Central Range system (DE VICENTE *et al.*, 1996). Also the small Calatayud - Daroca and Teruel basins and the peripheral Valles - Penedes and Lisbon basins were subject to high subsidence rates during the (Early) Langhian. Dextral strike-slip movements resulted in the development of some small, intra-montane basins, such as the Cerdanya and Seu d'Urgell basins in the eastern part of the Axial Pyrenees (ROCA, 1996). In all Iberian basins continental sedimentation persisted, which was characterised in the Ebro, Tagus, Duero, Calatayud - Daroca and Teruel basins by the accumulation of alluvial deposits along the basin margins, and of lacustrine successions in the more central parts of the basins. In the Ebro, Calatayud - Daroca and Tagus basins lacustrine sedimentation was mainly evaporitic (CALVO *et al.*, 1996; ORTI & SALVANY, 1997), whereas, as before, mainly calcareous sediments were deposited in the lacustrine environments of the Duero basin (MEDIAVILLA *et al.*, 1996). The sedimentary sequences of the peripheral basins in the north-east (e.g., Valles - Penedes; CABRERA & CALVET, 1996) and in areas in the west (Portugal; ANTUNES *et al.*, 1999) portray lateral transitions from continental to marine, littoral environments.

II.1.2.- Aquitaine basin

The areal distribution of Early Langhian marine depositional environments was relatively limited as compared with those in Early Miocene (Early Burdigalian) and late Middle Miocene (Serravallian) times, despite an earliest Middle Miocene sea level highstand (Hardenbol *et al.*, 1998). Langhian strata accumulated in two main areas, i.e. in the Saubrigues Gulf and the Central Aquitaine Gulf. In the Saubrigues Gulf (south-westernmost Aquitaine) predominantly marly sediments were deposited in littoral environments. These sediments filled up the downstream parts of the palaeocanyon which originated during the major eustatic lowstand at the Rupelian-Chattian transition (CAHUZAC *et al.*, 1995). In the Central Aquitaine Gulf (Gulf of Manciet - Baudignan) shallow-marine, littoral and, locally, lagoonal sediments accumulated. They consist of sands, shelly marls and crags. Locally, subreefal deposits with hermatypic and ahermatypic scleractinians were laid down (CAHUZAC & POIGNANT, 2000). In the easternmost part of the Aquitaine basin lacustrine, more or less molasse-type sediments were deposited. In the present-day offshore region of the south-eastern Bay of Biscay outer platform marly sands and limestones accumulated.

II.2.- Western Europe

II.2.1.- Alpine foreland basins

By the beginning of the Langhian, the continentalisation of the North Alpine Molasse basin, as initiated in its eastern part during the Late(st) Burdigalian, was abruptly completed all along the Eastern and Central Alps. Marine depositional conditions persisted only at the western extremity of the foreland basin, near the Rhone graben (Valence basin). The earlier established, generally ENE-WSW directed axial drainage pattern prevailed (BERGER 1996; SISSINGH, 1997). Uplift of the southern spur of the Bohemian massif led to a palaeogeographic disconnection with the Central Paratethys. Along the southern border of the basin, the Alpine thrust front continued to be active and alluvial fans continued to be fed with large quantities of gravel derived from the rising Alps (BACHMANN *et al.*, 1982). Adjacent alluvial plains were incised by meandering rivers and channels. A composite and repeatedly changing "Palaeo-Rhone" river system received water from rivers flowing towards the basin from the Alpine foreland in the north and from the Alpine orogenic wedge in the south. It discharged into a narrow sea which extended northwards through the Rhone graben from the Western Mediterranean basin (DEMARCQ & PERRIAUX, 1984; SISSINGH, 1997). Floral assemblages recovered from the Langhian - Early Tortonian molasse sequence indicate subtropical to temperate climate conditions.

II.2.2.- European rift systems

In the Rhine graben the Lower *Hydrobia* Beds conclude the Burdigalian depositional record. The Langhian depositional period of this basin is represented by the Upper *Hydrobia* Beds (BEST, 1975; ROTHAUSEN, 1988; SISSINGH, 1998). Both sedimentary sequences generally reflect brackish to marine environmental conditions of deposition. The youngest strata of the Langhian are more or less limnic. In the Lower Rhine embayment continental deposition prevailed in the larger part of the basin (HAGER & PRÜFERT, 1999). In this basin swampy

coastal lowlands occurred since the Rupelian - Chattian transition. They were generally confined to the southern part of the basin until the beginning of the Burdigalian, when the shoreline suddenly regressed in a north(west)ern direction and the coastal swamps concomitantly progradated in the same general direction. At the end of the Burdigalian, a eustatic lowering of the sea level induced a particularly pronounced and short-term northward progradation of the coastal lowland. Contemporaneously with a rapid and distant shift of the coastline and its adjoining marshes, the coeval marine regression culminated at that time with the development of an extended browncoal deposit, the Morken Seam. Following a major eustatic marine flooding over this seawards-enlarged coastal-lowland region, the terminal Langhian Frimmersdorf Seam developed under similar, regressive environmental and depositional circumstances. Marine communication between the Hessen depression and the Rhine graben did not exist anymore since the middle Chattian. From this time onwards, the eastern branch of the triple junction at the northern end of the Rhine graben became increasingly continental (MEIBURG & KAEVER, 1986). At the end of the Chattian, swamps re-appeared in the Hessen depression, as testified by lignites. During the Aquitanian and Burdigalian, fluvial sands were predominantly laid down in most parts of the Hessen depression. As far as known, the sea returned into the basin for the last time during the Langhian (from the north).

In the area of the Rhone graben, Langhian to Serravallian deposition was controlled by a S-N directed marine incursion into the terrestrial region of the central and northern parts of the rift (CAVELIER, 1984; SISSINGH, 1998). Tides occurred in this time-transgressively enlarged area of marine sedimentation, like in the preceding Burdigalian seaway. Littoral sands and conglomerates were most commonly deposited. Deltaic fans of conglomeratic material predominated in the Valence basin and near the Alpine thrust front, farther to the (north)east. In the contemporaneous Gulf of Lions *Avicenna* mangroves still occurred during the Langhian, in association with coral reefs. In the Massif Central intra-graben rivers continued to transport sandy erosion products to the north.

II.3.- Central Europe / Carpatho - Pannonian region

II.3.1.- General features

Subduction roll-back was the main process controlling the evolution of the Carpathian - Pannonian region during the early Middle Miocene. There was a sharp, overall decrease in average accumulation and subsidence rates in the Outer Carpathian (foredeep) basin systems, accompanied by a pronounced, lateral, eastward shift of foredeep depocentres, which was coeval with the opening of intra-Carpathian basins (MEULENKAMP *et al.*, 1996). The acceleration of the lateral foredeep depocentre migration may have been the response to the inception of slab detachment at about the Early - Middle Miocene transition, resulting in initial rifting of the overriding Alcapa and Tisza - Dacia microplates (CSONTOS, 1995) and, consequently, the opening of back-arc basins in association with major strike-slip displacements. The palaeogeographic reorganisation resulted in the development of mainly open-marine depositional environments and in the accumulation of predominantly clayey sequences, locally deposited at depths of up to about 600 m (e.g., BALDI, 1999). The larger invertebrate and foraminiferal associations indicate fairly uniform, tropical to subtropical climate conditions all over the arc - back-arc realm, from the west (Vienna basin) towards the south-east (Transylvanian basin, south-eastern Carpathians) during the Early Badenian. Volcanic activity played an important role, as evidenced, for instance, by the alternation of generally deep-water clays / marls and tuffites (with some conglomerate and sandstone interbeds) in the Transcarpathian, Transylvanian basin and in the Outer Carpathian foreland basins of Romania. These sequences belong to the molluscan *Neopycnodonta navicularis* Zone (MOISESCU & POPESCU, 1980). The upper beds belong to the planktonic foraminiferal *Candorbulina glomerosa glomerosa* Zone (POPESCU, 1970); their calcareous nannoplankton associations can be assigned to the NN4b (*Calcidiscus leptoporus*) and NN5a (*Geminithella rotula*) subzones (MARUNTEANU, 1992; 1998).

Marine corridors connected the Mediterranean realm with the basin systems of the Outer Carpathian foredeep and the intra-Carpathian domains of the Central Paratethys across Slovenia ("Trans-Tethyan-Trench-Corridor"), and, possibly, also in the south-east. However, the position of this presumably second connection has not been ascertained thus far (for a discussion, see RÖGL, 1998). Marine connections between the Outer and intra-Carpathian domains became severed in the course of the Middle Miocene. This caused the mid-Badenian (Wieliczian) salinity crisis, as reflected by the accumulation of evaporites, including halite.

II.3.2.- Regional aspects

II.3.2.1.- Outer Carpathians

In latest Early Miocene to earliest Middle Miocene times (Karpatian in terms of regional Central Paratethys stages) the north-vergent thrusting of the Eastern Alps ended, concomitantly with the acceleration of subduction roll-back in the Western and Eastern Carpathians (KOVAC & MARTON, 1998). Folding and thrusting in the external zones of the Outer Carpathians led to the development of a new, active thrust front formed by the Waschberg - Zdanice, Silesian - Subsilesian, Skole - Skiba, Tarcau, and Marginal Folds zones (see Fig. 17.3). SW-NE striking sinistral shear between the front of the Zdanice unit and the Magura nappe group was induced by the latest Early to early Middle Miocene (Karpatian to Early Badenian) oblique collision of the Western Carpathians with the Bohemian massif (FODOR *et al.*, 1995). These processes led to major contractions in the Magura, Silesian and Subsilesian zones, resulting in about 60% shortening (OSZCZYPKO & THOMAS, 1985; OSZCZYPKO & SLACZKA, 1989; OSZCZYPKO, 1997). In the central and southern parts of the Eastern Carpathians, early Middle Miocene (Badenian) tectonics, related to a ENE-WSW to E-W oriented compressional stress field (MATENCO, 1997), caused large-scale contractions in the

Tarcau and Marginal folds units. Areal differences in thickness of the sedimentary wedge can be linked with inferred variations in amounts of shortening.

II.3.2.2.- Intra-Carpathian region

The N-S oriented compression in the Eastern Alps was coupled with sinistral displacements along the Salzachtal - Ennstal and Mur - Murz - Leitha fault systems. According to FODOR (1995) and KOVAC *et al.*, (1997), strike-slip displacements also account for the opening of the pull-apart Vienna basin (straddling the Eastern Alps - Carpathian transition) in Karpatian time. NW - SE extension occurred in the western and central parts of the Carpathian - Pannonian region. Movements along N-S to NE-SW trending listric faults bordering the Eastern Alpine domain caused the opening of the Danube basin, which was associated with unroofing of core complexes (TARI *et al.* 1992; HORVATH, 1993; KOVAC *et al.*, 1994). Sedimentation in the South Slovakian - North Hungarian - Novohrad - Nograd basin (southern parts of the Western Carpathians) was controlled by movements along NW-SE oriented extensional faults (VASS *et al.*, 1993); the same holds true for the East Slovakian basin (north-western part of the Transcarpathian depression). A major phase of NE-SW stretching affected the central part of the Pannonian region, situated "on top of" the Tisza microplate (CSONTOS *et al.*, 1991). This Early Badenian tectonic trend continued into the domains of the Great Hungarian Plain, where the deepest depocentres developed (e.g., Hod, Mako and Bekes basins). According to HORVATH (1995), rifting came to a close after the middle Badenian compressional event, as inferred from the widespread unconformity observed in the Pannonian basin. Also in the Drava and Sava basins (westernmost part of the Tisza microplate and transitional area towards the Dinarides), subsidence and sedimentation were controlled by displacements along NW-SE directed normal faults (CSONTOS & HORVATH, 1995; PRELOGOVIC *et al.*, 1995). Farther to the east, major subsidence along similarly oriented faults caused the subsidence of the western parts of the Apuseni region, resulting in the opening of the Beius and Zarand basins (see also Gyorfi and Csontos, 1994). The opening of basins in the domains of the Tisza (- Dacia) microplate may have been associated with Karpatian to Early Badenian clockwise rotations (see also BALLA, 1984; KOVAC *et al.*, 1989; CSONTOS *et al.*, 1992, 1995; KOVAC *et al.*, 1994; CSONTOS & HORVATH, 1995; MORLEY, 1996). The overall stretching of the evolving back-arc region (CSONTOS & HORVATH, 1995) may have resulted in the formation of isolated core complexes, which, in turn, may account for the uplift of the Mecsek mountains and the Transdanubian Central Range in the west, and the Apuseni mountains in the east.

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

Anoxic conditions which had been characteristic for many of the deep-water parts of the Oligocene and Early Miocene basin systems, had ended in pre-Tschokrakian, i.e., Tarkhanian time. Contemporaneously, hitherto existing free exchange of water masses between the Western / Central Paratethys and the Eastern Paratethys came possibly to a close as well. This principal change in the general palaeogeographic configuration of the Paratethys is indicated by the similarity of Tarkhanian and Karpatian faunas and by the great differences between, for instance, the larger invertebrate (especially gastropods) and foraminiferal associations of the successive Tschokrakian and time-equivalent strata in respectively the Eastern and Central Paratethys realms. This palaeogeographic reorganisation may have been related to the Styrian orogenic phase, which caused uplift of the axial zone of the Eastern Paratethys i.e., the emergence of the Greater Caucasian archipelago. The resultant enhancement of the relief led to increased slope instability and subsequent deposition of coarse-grained terrigenous clastics, including turbidites (SCHERBA, 1993).

The Early Tschokrakian sequences are transgressive, particularly so in the northern and eastern parts of the basin system. The sea invaded, amongst others, the North Caspian and North Ustjurt embayments. Irrespective of the regional transgression, the connections with the world oceans became more restricted. This led to the development of endemic biotas by the end of the Early Tschokrakian. Restricted Early Tschokrakian exchange of water masses with the Mediterranean realm probably occurred via the Central Araksian Strait and through basins in Iran and eastern Turkey (GONTSHAROVA, 1989; ILJINA, 1993; GONTSHAROVA & SCHERBA, 1997). In addition, significant palaeogeographic modifications during the Tschokrakian are mirrored by an environmental change from a prevalence of relatively open marine conditions, albeit with fluctuating salinities, towards one of semi-marine environments inhabited by impoverished, euryhaline faunas (salinities less than 15 promille) in later Tschokrakian time. It seems that a land bridge connected the Greater Caucasian domain with Asia Minor and Africa by mid-Tschokrakian time. This development allowed the immigration of mammals (*Orycteropus*, *Kubanochaerus*) from Africa into the Pre-Caucasus domain (ZEGALLO, *in* GONTSHAROVA & SCHERBA, 1997). Throughout the Tschokrakian, climatic conditions were probably subtropical (GONTSHAROVA, 1989). Warm-water molluscs (*Pteria, Perna, Isognomon, Limaria, Chama, Alaba, Obtortio*), mesophylic subtropical floral elements (RAMISHVILI, 1983) and indications for the existence of mangrove-type vegetation near the eastern Pre-Caucasus islands, as well as the composition of the insect fauna, suggest that the "second Miocene climatic optimum" (AKHMETIEV, 1993) persisted from the Tarkhanian into the Tschokrakian. Relatively arid conditions existed in the Transcaspian domain, as evidenced by the vast distribution of evaporites and reddish-coloured clastics.

II.4.2.- Regional aspects

II.4.2.1.- Eastern European platform and Scythian plate

The Eastern European platform was an emerged, continental realm onto which the Ukrainian massif, the Donetz land and the Ural High were the main positive topographic features in the Tschokrakian. The Russian Lowland was drained by the Palaeo-Don, whereas the Palaeo-Donetz drained the eastern Dniepr - Donetz depression. The vast Palaeo-Don delta was periodically flooded by the sea (J.I. IOSIFOVA & A. ZASTROZHNOV, pers. comm.). Principal areas of fluvio-lacustrine sedimentation were the Dniepr-Donetz and Precaspian depressions. During the Early Tschokrakian the marine North Caspian Gulf extended as far as the eastern part of the Precaspian depression. The Scythian plate corresponded to the northern shelf area of the Eastern Paratethys. Shallow-water, sandy to silty and biogenic sedimentation predominated. The West Kuban and Terek - Caspian depressions were subsiding and filled with thick sequences of calcareous muds with fluxoturbidite intercalations in the former, and with silts and sands in the latter basin. Algal and bryozoan reefs developed on the top of escarpments.

II.4.2.2.- Turanian plate and Kopet Dagh

Like the shelfal Scythian plate, the Turanian domain was also subjected to relief differentiation in Tschokrakian time. Besides the localised North Ustjurtian and Fore-Kopet Dagh depressions, which existed already since the Tarkhanian, the sea invaded other newly-formed regional depressions, such as those of the North Caspian and South Mangyshlak regions and also depressions of the South Aral - Kopet Dagh (Assakeaudan, Upper Uzboy, Ilym - Balkuin) regions. Some of these were periodically transformed into (semi-)closed lagoons, in which evaporites accumulated. Interfingering of marine clastics and reddish continental successions occurred in the eastern parts of the North Caspian and North Ustjurtian gulfs. Along the margins of the marine basins various types of continental deposits accumulated, which, however, cannot be dated precisely. Reddish, silty to clayey molasse-type successions were deposited in the eastern Kopet Dagh and Fore-Kopet Dagh depressions (Karagaudan Suite) and in the Fore-Tien Shan and Tadjic depressions (Hengou Suite; east of mapped area). Middle Miocene proluvial and lacustrine deposits, comprising reddish clays with gypsum intercalations, were laid down in Kyzylkum. Greyish, lacustrine and fluvio-lacustrine clays and silty sands were deposited in the Turgaj and Teniz depressions of the Kazakhstan Highland and in the West Siberian Lowland areas.

II.4.2.3.- Black Sea depressions and Caucasus

Deep-water environments persisted in the Western Black Sea, the Eastern Black Sea and the South Caspian depressions. Possibly, the latter depression extended into the south-eastern Caucasus (Kobystan trough), where fine-grained terrigenous clastics rich in organic matter witness of anoxic conditions. The Western and Central Caucasus evolved into a land massif, which was temporarily connected with the Lesser Caucasus, probably via the Dzirul massif. This continental connection has been inferred from the composition of Pre-Caucasus mammal associations (Belomechetka). Indications for the presence of mangroves and the nature of leaf remains (AKHMETIEV, 1995), insect assemblages and lagoonal ostracode associations suggest altogether the previous existence of a large Western - Central Caucasian island and an archipelago farther to the east. The uplift of the Lesser Caucasus, Talysh and Elburz chains continued in Tschokrakian time. Accumulation of sandy and silty sediments prevailed in the southern half of the Transcaucasian southern shelf domain. The Adzharo - Trialet depression (south-eastern part of the Eastern Black Sea depression) was reduced in size. Its southern part was filled with terrigenous clastics, which originated from the Lesser Caucasus. Apparently, accumulation of sandy muds kept pace with subsidence in the "second order" Maykopian depressions of Talysh, Tuapse and Shemakha in the course of the later part of the Middle Miocene.

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi - Oman domains of the Arabian peninsula

In latest Early Miocene times, Alpine orogenic activity induced folding and uplift of the Oman mountains (WYNS *et al.*, 1992; LE MÉTOUR *et al.*, 1995). During the Late Burdigalian to Early Langhian, parts of the Arabian platform (which were emerged since the Middle Eocene) were subject to a major transgression (CAVELIER *et al.*, 1993). The transgression came from the north (present-day Arabian Gulf). The sea flooded the area of Rub' al Khali, which was bordered to the north-east and the east by an Eo-alpine chain and the Haushi - Huqf structural high. Very shallow-marine, locally restricted environments of deposition, represented by the Dam Formation, passed to the west into a lacustrine depositional setting, in which carbonates (Dawkah Formation) were laid down in north-western Dhofar. North of the Haushi - Huqf high, as far as the region of Umm as Samin, the marine deposits passed into lagoonal sediments with marine interbeds indicating periodical incursions of the sea. These depositional conditions were associated with the accumulation of evaporites and terrigenous clastics. Shallow-marine, reefal and clastic deposits accumulated on a narrow platform to the north of the Oman mountains (WYNS *et al.*, 1992). In southern Dhofar, the youngest part of the turbiditic Mughsayl Formation was deposited along the foot of the slope of the Aden rift at the beginning of the Langhian. Farther to the east, along the margin of the Arabian peninsula, the deposition of prograding carbonate prisms continued throughout the Early Langhian (fore-reef limestones of the Warak Formation; PLATEL *et al.*, 1992c).

A major change of the tectono-sedimentary regime occurred in the course of the Langhian. This change was associated with tectonic tilting of the Arabian plate to the south-east. Concomitantly the sea withdrew completely from the hitherto existing marine depositional area of the

Dam Formation. Marine deposition was succeeded by the development of palaeosoils (BERTHIAUX & PLATEL, 1992). From the Late Langhian to the Serravallian, a "final", major transgression from the Indian Ocean affected all of the north-eastern Dhofar and central Oman region (PLATEL *et al.*, 1992a). This led to the development of very shallow shelf environments, in which the Ghubbarah Formation was deposited. This sequence portrays a palaeogeographic setting of a gulf with reefs, which passed to the west into restricted lagoons. Along the margin of the Rub' al Khali, this 200 km wide gulf was bordered by a fairly continuous belt of lacustrine environments with carbonate deposition, as represented by the Montasar and Ghaba formations. Along the foothills of the contemporaneous Oman mountain chain, predominantly fluviatile and some lacustrine sediments accumulated, which are represented by the widely-distributed Barzaman Formation (ROGER et al, 1992a; WYNS *et al.*, 1992). In eastern Arabia, the Langhian regression led to the deposition of the continental siliciclastic Hofuf Formation (POWERS *et al.*, 1966), which has been dated as Late Langhian to Early Serravallian by means of vertebrate associations (SEN & THOMAS, 1979). By the end of the Langhian, the coastal belt of the Gulf of Aden was affected by a major tectonic uplift and rejuvenation of major listric fault systems, which led to the development of cliffs along the Salalah coastal plain, southern Oman (PLATEL & ROGER, 1989; PLATEL *et al.*, 1992b; ROGER *et al.*, 1992b). For the first time in its existence, the adjacent narrow shelf received conglomerates, which occur interstratified with carbonates (Adawnib Formation). More or less contemporaneously, littoral and, successively, continental, molasse-type sediments belonging to the Salmiyah Formation were deposited along the northern slopes of the Oman mountains (WYNS *et al.*, 1992).

II.5.2.- Israel

Reefal limestones developed in the area of the present-day coastal plain of Israel and in the Shefela region (Ziqlag Formation). They reflect the (relative) rise of sea level (BUCHBINDER, 1996a and b; BUCHBINDER & ZILBERMAN, 1997) in Langhian time. Most of the domain of the Negev, Judea and Galilee regions was emerged. Sinistral displacements along the Dead Sea Transform fault continued; these movements were coupled with shortening in the Palmyrides.

II.5.3.- Egypt

The Middle Miocene sequences are separated by an unconformity from those of the Early Miocene (SAID, 1990). Their vertical as well as lateral facies developments across the various tectonic domains differed strongly. In the northern Western Desert the Langhian is represented by shallow-marine limestones constituting the lower part of the Marmarica Formation, which capped the northern plateau. North of the Qattara depression (north-western Egypt) the limestones reach a thickness of about 1000 m. In the Nile delta region the Langhian is represented by bathyal clays and shales composing the upper part of the Sidi Salem Formation. Time-equivalent shallow-marine carbonates and shales of the Genefe and Hommath formations are exposed in the Cairo - Suez area. Deep(er)-water marine shales and marls (Upper Rudeis Formation) were deposited in the Gulf of Suez rift during the Early Langhian (QUDA & MASOUD, 1993). The unconformity between the Lower and Upper Rudeis formations is attributed to a major tectonic event straddling the Early - Middle Miocene transition. In the marginal parts of the northern Red Sea rift the Langhian is represented by coralline limestones and evaporitic calcareous deposits of the Um Mahara Formation (PURSER *et al.*, 1993).

II.5.4.- Tunisia

Langhian depositional domains occupied a large portion of the Tunisian platform, including most of central and north-eastern Tunisia. They corresponded in particular to NW-SE and NE-SW trending (sub)basins (YAICH, 1992). In central Tunisia these grabens and half-grabens were inherited from the Oligocene (YAICH, 1997). Additional, W-E (central Tunisia; YAICH, 1984; TURKI, 1985; TURKI *et al.*, 1988; BLONDEL, 1991) and NW-SE trending extension (PHILIP *et al.*, 1986) resulted in a multidirectional rift system controlling deposition in Langhian times. The sedimentary record indicates two major episodes of (relative) rise of sea level. One occurred in the Early Langhian (*Praeorbulina glomerosa*, Zone; HOOYBERGHS, 1973; HOOYBERGHS & YAICH, 1996): the other in Late Langhian times (*Orbulina suturalis* Zone; HOOYBERGHS, 1973, 1977; HOOYBERGHS & GHALI, 1990). The Langhian successions overlie Cretaceous to Early Miocene deposits (YAICH, 1992). Those associated with the Early Langhian transgression are composed of various calcareous and clayey sequences. They attained greatest thicknesses in depressions which either existed at least from the Early Miocene onward, or originated in response to Langhian synsedimentary faulting. At various places glauconitic clastics and/or polymict bioclastic limestones (BLONDEL, 1991) occur at the base of the Langhian (Oued Hammam Formation; HOOYBERGHS, 1973). These basal strata indicate deposition in environments ranging from coastal-marine to deep-marine.

In Early Langhian times, the western and south-western parts of Tunisia, as well as the southern domains including parts of the Tunisian Sahara, were emerged. The Early Langhian successions are overlain by (sandy) calcareous sediments, which are often conglomeratic at their base (Ain Grab Formation; BUROLLET, 1956). These sediments, which have a remarkably wide and continuous distribution all over the Tunisian platform, reflect the second, Late Langhian transgression. The ensuing sedimentation was characterised by the accumulation of a composite of siliciclastic and calcareous sequences in shallow-marine platform environments, which passed into deeper marine environments towards the east. The shallow-marine deposits cover large parts of central and northern Tunisia, whereas coastal-marine siliciclastic successions are confined to the areas bordering the "North-South Axis" in central Tunisia. The deepest-marine environments of the Late Langhian were located in the area of Cap Bon (BEN ISMAIL-LATTRACHE, 1981; BLONDEL, 1991, YAICH *et al.*, 1994) and in the Gulf of Gabes (YAICH, 1997).

22.- LATE TORTONIAN (8.4 - 7.2 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

The Late Tortonian ranges from about 8.4 to 7.2 Ma (Fig.17.1). In terms of marine microplankton biochronology it covers the early part of the calcareous nannoplankton *Discoaster quinqueramus* zone (NN11) and the upper part of the Mediterranean planktonic foraminiferal *Neogloboquadrina acostaensis* zone. In the terrestrial record the Late Tortonian interval corresponds to part of the Turolian (micromammal zones MN11 and MN12). The position of regional stages of the Central and Eastern Paratethys relative to the Late Tortonian is less unambiguous in detail. The Late Pannonian of the Central Paratethys may be considered the equivalent of the Late Tortonian. Late Maeotian successions have been included in the maps for the Eastern Paratethys, based on radiometric dating of the Late Maeotian (~ 8-7 Ma; CHUMAKOV, 1993).

I.2.- Structural setting and kinematics

A major tectonic reorganisation affected large parts of the African - Eurasian collision zone in the course of the Tortonian. In the Mediterranean area this resulted, for instance, in the opening of the Tyrrhenian and southern Aegean back-arc basins at the beginning of the Late Tortonian. In the Central Paratethys the end of lateral (i.e., arc-parallel) foreland depocentre migration coincided approximately with the inception of thermo-tectonic subsidence in the Great Hungarian Plain (Pannonian) back-arc basin system during the early Late Miocene. Tectonic reorganisation around the Middle - Late Miocene transition also caused large-scale palaeogeographic changes in the Eastern Paratethys, including the disrupture of marine connections between the Euxinian and Caspian basins across the Transcaucasus area and the origin of the Greater Caucasus as a mountain range. The Late Miocene tectonic events had pronounced counterparts in the Peri-Tethyan domains proper, expressed, for instance, by strike-slip faulting and reactivation of pre-existing fault systems on the Iberian massif. On the southern Peri-Tethys platform the Dead Sea Transform evolved into the Dead Sea Rift.

I.3.- Outlines of palaeogeography and palaeoenvironments

In the Late Tortonian, large parts of the northern Peri-Tethys platform were emerged and formerly comprehensive sedimentation realms started to break up in response to overall uplift, particularly so in the western and central domains. As compared to the Langhian, marine influences had strongly decreased. On the Iberian peninsula sedimentation continued to be mainly confined to the accumulation of lake deposits in intra-montane basins and to the deposition of marine sediments in various basins in the south, including the Betic corridor. In various basins of the Cainozoic rift system in Western Europe, sedimentation had ended or become confined to the accumulation of exclusively continental (predominantly fluvial) deposits. The comprehensive NW-SE oriented basin system ranging from the North Sea to the Black Sea (via the Polish and Pripyat - Dnieprovian Lowland basins), which still existed in the Early Langhian, had been broken up. In the Tethyan - Peri-Tethyan transitional domains of Central Europe the main environmental changes relative to the Early Langhian include the tectonically-induced termination of sedimentation in the (overfilled) peri-Alpine molasse basins and the disappearance of marine connections between the Mediterranean basins and the intra-Carpathian back-arc basins, in which only fluvio-lacustrine to brackish sediments accumulated from the Middle - Late Miocene transition onward.

In the east, the Late Miocene palaeogeography and environmental conditions differed strongly from those before. The outlines of the present-day Black Sea depression originated at the time, while marine connections between the Euxinian and Caspian basins via the Transcaucasus area no longer existed and the sea had withdrawn from the Turan domain. Some peri-Carpathian

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basins, which hitherto formed part of the Central Paratethys, were palaeogeographically and environmentally incorporated in the Eastern Paratethys domain in the Late Miocene (transition Sarmatian - Maeotian). These domains were characterised by semi-marine to brackish conditions. Time and again, however, ephemeral incursions from the Mediterranean via the North Aegean corridor resulted in marine conditions, as evidenced by the occurrence of marine microfossils (calcareous nannoplankton) and macrofossils.

The most important changes in palaeogeography and environmental setting on the southern Peri-Tethys platform relative to the Early Langhian include the withdrawal of the sea from the eastern part of the Arabian platform. The marine corridor between the Mediterranean Sea and the Red Sea across the Suez sill was closed prior to the Late Tortonian. In Late Miocene time, the Gulf of Suez - Red Sea rift system was only connected with the Indian Ocean. Sedimentation in the rift system had become characterised by the accumulation of thick sequences of evaporites, including halite, from about the late Middle Miocene (Serravallian) onwards.

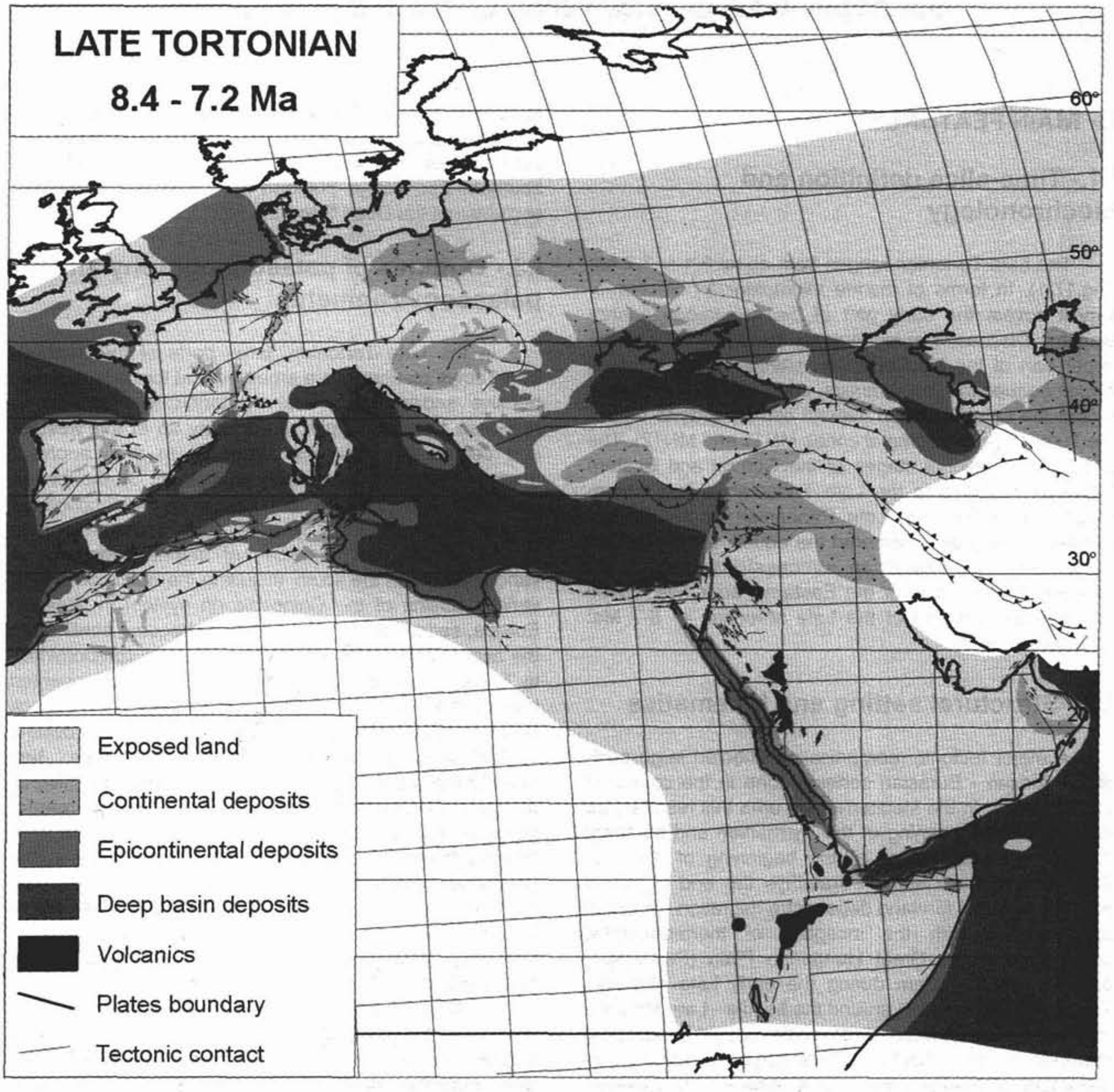

Fig. 21.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Late Tortonian.

II.- Description of Domains

II.1.- South-western Europe

II.1.1.- Iberian Peninsula

The overall, N-S oriented compression associated with the convergence of the African and Eurasian plates persisted in the Late Tortonian. This resulted both in uplift and in displacements along strike-slip faults, especially in

the south-eastern parts of Iberia where various basins developed (SANZ DE GALDEANO & VERA, 1992). A major palaeogeographic - palaeotectonic change (basically related to radial extension of the Iberian range) affected the interior continental basins of Iberia, as reflected by the reactivation of pre-existing faults (including the Plasencia fault), which resulted in displacements along N-S, NW-SE and NE-SW trending faults (DE VICENTE *et al.*, 1996). Relative to the Early Langhian, the dimensions of depositional areas in the interior basins were reduced. In the Ebro basin only locally alluvial deposits and lacustrine carbonates accumulated in Late Tortonian (Early to Middle Turolian) time. A similar decrease in size of depositional areas occurred in the Duero (MEDIAVILLA *et al.*, 1996) and Tagus basins. In these basins continental sedimentation was confined to the central parts, where shallow lake carbonates accumulated adjacent to the distal parts of N-S trending fluviatile systems (CALVO *et al.*, 1996). In all three major interior basins a change had taken place from inner (endorheic) to external (exorheic) drainage conditions. Other small interior basins (e.g., those of Calatayud - Daroca and Teruel and the external Betics) were filled up with thick successions of alternating alluvial and (calcareous) lake deposits (CALVO *et al.*, 1993). In the peripheral, small basins bordering the Mediterranean, generally thick successions of alluvial to lacustrine sediments were deposited, but deposition in shallow to deeper-marine environments prevailed in the most external basins.

II.1.2.- Aquitaine basin

During the Late Tortonian, marine conditions only prevailed in the westernmost part of the basin, where sandy marls accumulated in inner and outer shelf environments of deposition. The westward progradation of the shelf contributed to the infill of the oceanic domains to the west. Neritic, sandy marls were deposited along the south-eastern margin of the Bay of Biscay (DURAND, 1974). Molasse-type sediments were laid down in the eastern(most) parts of the basin and along the northern front of the Pyrenean orogen (CROUZEL, 1956). Farther to the north-west, pelagic deep-water sedimentation continued (DSDP Leg 48; MONTADERT & ROBERTS, 1979).

II.2.- Western Europe

II.2.1.- Alpine foreland basins

Alluvial fan and plain facies continued to predominate in the North Alpine Molasse basin until the Tortonian. Since its youngest known deposits are of Early Tortonian age, the Late Miocene to Pliocene development of the North Alpine Molasse basin is poorly known. At the westernmost end of the basin, Langhian and Serravallian fan-delta sands and conglomerates ultimately occluded the western, tide-influenced entrance of the basin. Wholesale uplift, causing regional erosion in the northern foreland basin (LEMCKE, 1974) and coinciding with rapid elevation of the Alps (SCHMID *et al.*, 1996), apparently started subsequently during the Tortonian. It induced the termination of molasse deposition and the transformation of the North Alpine Molasse basin into an area of largely non-deposition. Initially, fluvial transport of clastic erosion products derived from the rising Alps was directed to the west, as in the preceding period. However, basin uplift was strongest in the west. In response to the resulting tilting of the basin, various rivers draining the proximal Alpine foreland started to flow increasingly eastwards from the latest Miocene onwards. The basinal uplifting may be related to isostatic adjustment of the lithosphere in response to gradually waning thrust movements and foreland loading by the Apulian plate, in association with unflexing of the overthrusted European plate.

II.2.2.- European rift system

During the Serravallian, deposition in the Rhine graben proceeded in non-calcareous, largely lacustrine facies. Minor streams draining bounding highs seem to have deposited sandy clastics in the graben. Following a period of non-deposition/erosion during the Early Tortonian, clastic sedimentation re-occurred in fluvial facies during the Tortonian and Messinian. The deposits represent for the greater part the intra-graben river Rhine, which by then flowed northwards from the Vosges across the Rhenish Massif to the Lower Rhine embayment and North Sea basin. The predominance of fluviatile deposition after a period of regional uplift and erosion is attributed to ongoing widespread tectonic activity during the Early Tortonian (SISSINGH, 1998).

In the Rhone graben the sea regressed eustatically close at the Serravallian-Tortonian transition. As a result, brackish-water and fluvio-lacustrine, mainly clastic sediments were deposited in the northern section of the graben and more conglomeratic material was laid down in the central graben segment, where littoral marine conditions continued to exist until the Early Tortonian. Depositional conditions became increasingly continental during the Tortonian. The concomitant southward retreat of the sea is particularly reflected by the accumulation and progradation of conglomerates and lignite-bearing strata. These sedimentary changes may be related to a general uplift of the Rhone graben area rather than to a persistent drop in global sea level (CAVELIER, 1984; SISSINGH, 1998). The regional elevation may have included the Massif Central, where a large fluvial system continued to drain the old high towards the north. The Valensole basin became increasingly discrete and continental during the Late Tortonian. The general continentalisation of the Rhone graben and other, surrounding basins climaxed in the Messinian, when erosion and non-deposition became suddenly dominant over sedimentation, in relation to the terminal Miocene desiccation of the Mediterranean basin. A widespread and intensely erosive palaeohydrographic system of overall southward discharging rivers and minor streams was established in response to the Mediterranean salinity crisis during the Late Messinian (BALLESIO, 1972). It resulted, for instance, in the origin of the Rhone as the major river draining the region of the western North Alpine Molasse basin and the adjacent Alps, as well as the West Alpine foreland area. These pronounced changes in palaeogeography coincided with a terminal Miocene phase of increased tectonic activity, which basically affected all of peri-orogenic eastern France. Tectonic activity during the latest Miocene included in particular westward-directed

thrusting of the Jura over the eastern margin of the Bresse graben and a general re-emplacement of the Subalpine chains. At the same time, the Valensole basin became an important land-locked locus of detrital deposition.

II.3.- Central Europe / Carpatho-Pannonian region

II.3.1.- General features

As in Middle Miocene times, the Late Miocene geodynamic evolution of the Carpathian - Pannonian system was mainly controlled by subduction roll-back. The continuing subduction retreat in the Eastern Carpathians was probably associated with a new, early Late Miocene (Early Pannonian) episode of rifting, which was followed by the inception of thermal subsidence in the back-arc area (LANKREIJER, 1998). In contrast to the Early Langhian, when open-marine conditions prevailed in arc as well as intra-arc domains, brackish to fluvio-lacustrine environments were characteristic for the Central Paratethys in the Late Tortonian. Marine corridor(s) connecting the intra-arc domains with the Mediterranean no longer existed. In the Outer Carpathian domains ephemeral marine ingressions were confined to the Dacic basin (south-eastern part of the Outer Carpathians). The corresponding sediments of the "Middle" Maeotian (Late Oltenian to Early Moldavian) contain calcareous nannoplankton associations (with *Discoaster quinqueramus, Amaurolithus primus, A. delicatus*) indicative of the lower part of zone NN11 (FORNACIARI *et al.*, 1997; MARUNTEANAU & PAPAIANOPOL, 1998). Biogeographic data indicate that, in fact, the Dacic basin had become part of the Eastern Paratethys ("western embayment" of the Euxinian / Black Sea realm) in Late Tortonian time. The "Middle" Maeotian sediments of the Dacic basin are correlative to the Late Pannonian sequences of the Great Hungarian Plain basin system and the Transylvanian basin, in which fluvio-lacustrine to brackish-water deposits accumulated. Brackish conditions in the isolated intra-arc basins of the Central Paratethys are evidenced by molluscs and endemic nannoplankton assemblages belonging to the *Noelaerhabdus bonagali* zone (MARUNTEANU, 1997).

II.3.2.- Regional aspects

II.3.2.1.- Outer Carpathians

Late(st) Miocene thrust tectonics are only known to have occurred in the frontal parts of the transitional area between the Eastern and Southern Carpathians. Here, flexural downbending of the platform led to subsidence and sedimentation in the Eastern Carpathian foredeep and in the Getic depression, in which Late Miocene molasse-type deposits reach thicknesses of 1000 to 2000 m (DICEA, 1996). The active front of the Eastern Carpathians comprised the Subcarpathian Folded Neogene unit (Fig. 17.3), which was thrusted over the foredeep during the intra-Pannonian, Moldavian tectonic phase. In the Subcarpathian zone a major reduction of depositional areas occurred at about the Maeotian - Pontian transition (LINZER, 1996; MATENCO, 1997). Regional compressional stresses were oriented NE-SW to E-W in the Eastern Carpathians, and NW-SE to N-S in the Southern Carpathians and Getic depression (LINZER, 1996; MATENCO, 1997).

II.3.2.2.- Intra-Carpathian region

The Late Miocene evolution of the Pannonian basin system followed upon an episode of regional shallowing and local uplift around the Sarmatian - Pannonian transition (MEULENKAMP *et al.*, 1996). According to LANKREIJER (1998) a new episode of rifting occurred in the earliest Late Miocene (Early Pannonian) under influence of NW-SE extension (CSONTOS *et al.*, 1991; CSONTOS & HORVATH, 1995). At the time, subsidence of basin depocentres was probably controlled by left-lateral transtensional wrenching and by displacements along major listric faults, which offset the basement (CSONTOS & HORVATH, 1995). The Early Pannonian rifting episode was succeeded by a phase of widespread extensive, overall post-rift thermal subsidence (HORVATH, 1993; CSONTOS & HORVATH, 1995; MEULENKAMP *et al.*, 1996).

During the Pannonian, maximum subsidence in the western part of the back-arc basin system took place in the central parts of the Danube basin (Gabcikovo and Komjatice depressions), in which 2000 m of deltaic and lake deposits accumulated (TANACS & RADICH, 1991; MATTIC *et al.*, 1996; KOVAC *et al.*, 1997). Subsidence and sedimentation were controlled by NW-SE extension associated with displacements along NE-SW to NNE-SSW oriented faults (FODOR, 1995; FODOR *et al.*, 1995; HRUSECKY *et al.*, 1996; KOVAC *et al.*, 1997). Also in the Vienna basin NW-SE extension prevailed. The reactivation of NE-SW directed strike-slip faults may be considered the response to W-E compression in the domains of the Eastern Alps (DECKER & PERESSON, 1996). Strike-slip displacements along the mid-Hungarian lineament zone (see Fig. 17.3) compensated for the extension of the Tisza microplate, which, in turn, mirrors the subduction retreat in front of the Eastern Carpathians. These strike-slip displacements led to the formation of deep basins between the Kiskun basin in the west and the Derecke basin in the east. In the central parts of the subsiding zone (Zagyva, Jaszag and Hernad basins) deltaic sequences accumulated, with a thickness of up to 2000 m in the Jaszag basin (JUHASZ, 1991). During the subsequent post-rift phase of thermal subsidence, also depocentres of the Great Hungarian Plain basin (Hod, Mako and Bekes basins) received large amounts of clastics from the north-west, as evidenced by the progradation of deltaic sequences (up to 3000 m; JUHASZ, 1991).

The post-rift phase of overall thermal subsidence was temporarily interrupted by compressional tectonics associated with the intra-Pannonian, Moldavian orogenic phase (MATENCO, 1997), as evidenced by thrusting and / or folding in the Sava area, the area south of the Mecsek mountains and the region along the southern margins of the Transdanubian Central Range/mid-Hungarian lineament (CSONTOS *et al.*, 1991; CSONTOS & HORVATH, 1995) and in the Caransebes - Mehadia basin of Romania (MARUNTEANU *et al.*, 1995). The thrusting and folding ultimately resulted in the subsequent Pliocene inversion, which was comparable to the first inversion episode around the Sarmatian - Pannonian transition.

The Late Miocene extension was associated with volcanism in various parts of the intra-Carpathian region. Pannonian - Pontian calc-alkaline and alkali-basaltic volcanism is known from the volcanic chain of the Calimani - Gurghiu - Hargita mountains in the Transylvanian basin and the areas of Slanske Vrchy, Vihorlat and Gutii; basaltic volcanism from the Styrian and Danube basins (SZABÓ *et al.*, 1992; LINZER, 1996).

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

Tectonic movements related to the "Attic orogenesis" resulted in large-scale palaeogeographic reorganisations in latest Middle to early Late Miocene time. These reorganisations included the transformation of the Greater Caucasian region into a mountain chain and a major provenance area of terrigenous clastics from the Late Sarmatian - Maeotian onward. Foundering of the hitherto emerged Shatsky and Andrusov ridges resulted into a Black Sea depression with features more or less similar to those of the present-day. These changes in terrestrial topography and basin configuration were accompanied by the disrupture of the marine communication between the Euxinian and Caspian basins across the Transcaucasus area. This major change in palaeogeography resulted in the transformation of the Greater Caucasus into a peninsula connected to the Lesser Caucasus, and in the withdrawal of the sea from almost the entire Turan plate area. In the west, the Dacic basin became part of the Eastern Paratethys realm and the Maeotian basin system thus extended from eastern Serbia to south-western Ustjurt. However, the overall size of the basin system reduced relative to that during the Sarmatian. The Euxinian and Caspian basins were still connected via a strait north of the Stavropolian highland.

In the Early Maeotian, hemimarine environmental conditions prevailed in the Eastern Paratethys and the connections with the open-marine Mediterranean basin (probably through the north-eastern Aegean region, or across Turkey and Iran) were restricted. Eventually, the impoverished, euryhaline fauna of the Early Maeotian became extinct. In Late Maeotian time, brackish *Congeria*, *Theodoxus, Pseudamnicola* and *Turricaspia* associations were widely distributed. Ephemeral marine ingressions occurred, as evidenced by intercalations with *Mactra superstes* and *Sphaeronassa* and marine diatoms (*Asterolampha marylanica, Coscinodiscus asteromphalus, C. levisahus*). In addition, marine nannoplankton has been recovered from sequences of the Kerch and Taman peninsulas (LULIYEVA, *in* SEMENENKO, 1987) and the Apsheron peninsula. Palynological and mammal data from the Black Sea region are indicative of open, sparsely forested hinterlands with an arid climate similar to that of today's steppes or semi-deserts (NEOGENE SYSTEM, 1986).

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

The larger part of the East European platform corresponded to the Russian Highland, which was bordered by the Ural High in the east. The Ural High was a low, mountainous area flanked by hilly reliefs. Major "loci of deposition" probably occurred in the Pripyat - Dnieprovian lowland basin and the Precaspian lowland, although the Maeotian age of the continental deposits in the latter area has not been ascertained. Alluvial sands with remains of *Mastodon borsoni* and fresh-water fishes are known from the Donbass region (Janov Suite, Palaeo-Donetz deposits). The lower Ergeni Sands of the Palaeo-Don depositional system are of Maeotian Age (based on palynological data; J.I. IOSIFOVA and A. ZASTROZHNOV, pers. comm.). Vast alluvial plains occurred in the Palaeo-Dniepr, South Bug and Dniestr areas. Almost the entire domain of the Scythian plate was occupied by brackish, shallow seas in which calcareous, sandy and clayey deposits with abundant representatives of *Congeria* faunas accumulated. Deeper-water clays deposited in outer shelf environments are known from the Indol - Kuban depression only. The major positive structure of the Scythian plate was constituted by the Stavropolian Highland. The Indol - Kuban and Terek - Caspian Fore-Caucasian depressions originated along the concurrently evolving Greater Caucasian mountain range. They received large amounts of coarse clastics from this mountain chain, which were deposited in both marine and terrestrial environments. The sediments contain *Mastodon* and debris of fresh-water molluscs.

II.4.2.2.- Turanian plate, Tien Shan and Kopet Dagh

With the exception of the South Mangyshlak depression, deposition in the Turan domains occurred exclusively in continental facies during the Late Maeotian. The Kopet Dagh, Pamir, South Tien Shan, North Tien Shan and Karatau regions (partly east of mapped area) had evolved into mountain ranges. Coarse clastics accumulated in their peripheral molasse belts. Finer-grained, sandy to clayey lake and alluvial sediments were deposited to the north and north-east, i.e. in the Turan Lowland area and the Tadjic and Fergana depressions (east of mapped area). Areal differences in thickness of the continental sequences were controlled by movements along the Amudarja and Western Aral regional fault system. Late Maeotian accumulation of shallow-water shelf sediments was confined to the South Mangyshlak Gulf.

II.4.2.3.- Black Sea depression and Caucasus

The present-day contours of the Black Sea depression originated in the Maeotian. The north-western parts of the continental slope of the basin were covered by deltaic deposits supplied by the Palaeo-Danube and Palaeo-Dniestr river systems. These Maeotian sequences reach a thickness of 1100 metres. The Adzharo - Trialetic depression (with the exception of its north-western, Gurian region), evolved into a highland area merged with the Lesser Caucasus. In the adjacent southern shelf areas only two small embayments remained, the Riony bay in the west and the Kura depression in the east. The latter was transformed into an intra-montane continental basin, which fill is characterised by thick successions of conglomerates, sands and clays (Dusheti and Shirak

Suites) containing remains of terrestrial mammals and floras. In the deep South Caspian depression the so-called "Diatom Suite" was deposited. The depression extended as far as the Apsheron peninsula. Sandy, coal-bearing deposits were laid down in the shallow-water environments of the Kura depression.

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi-Oman domains of the Arabian peninsula

The major regression, which affected the domains of Saudi Arabia and interior Oman in latest Middle Miocene (latest Serravallian) times, resulted probably from the combined effects of an eustatic fall in sea level and a general tectonic uplift of the eastern part of the Arabian peninsula. Thick successions of fine-grained clayey sands and gravel, deposited in floodplain and (subordinate) lacustrine and paludine environments (Marsawdad Formation; BERTHIAUX & PLATEL, 1992), accumulated in the Rub' al Khali region. These deposits reflect the infilling of the vast Rub' al Khali basin through the progradation of distal clastic sequences derived from the Arabian - Yemen shield. They extend to the south as far as the northern Dhofar region and to the north into Qatar (Hofuf Formation *p.p.*; POWERS *et al.*, 1966; CAVELIER, 1970). In the Late Tortonian, Arabia and the western parts of interior Oman had become subjected to a long-lasting period of emersion. Deposition of piedmont conglomerates persisted along the Oman chain (Barzaman Formation; WYNS *et al.*, 1992), in part concomitantly with the accumulation of conglomerates north of the Dhofar mountains (Shisr Formation; CHEVREL *et al.*, 1992). In Late Tortonian to Pliocene times, thick piedmont conglomerates were deposited in coastal plain areas (Al Nar Formation in the Salalah plain, southern Oman; conglomerates in the Batinah plain of northern Oman). The eolian deposits (dunes; known as "aeolianites") of the Wahibah Sands of eastern Oman, which are supposedly Miocene - Pliocene in age, represent the oldest sediments evidencing the existence of desert environments in this region (GARDNER, 1988; LE MÉTOUR *et al.*, 1992).

II.5.2.- Israel

The Levantine domain was affected by large-scale uplift of the mountainous backbone and by the collapse of the "polygons" of the Dead Sea Transform fault system, which resulted in the origin and opening of the Dead Sea Rift basin. The latter process and the concomitant reactivation of pre-existing faults which opened the Yezreel Valley, enabled marine inundation of the Dead Sea Rift basin from the Mediterranean. This, in turn, resulted in the deposition of the evaporites (mainly halite) of the Bira - Zema Formation. Parts of the accomodation space of the Dead Sea Rift basin were filled by basaltic extrusives. In response to the Tortonian rise of sea level reefs (Pattish Formation) originated in the domain of the present coastal plain of Israel (BUCHBINDER, 1996b; BUCHBINDER & ZILBERMAN, 1997).

II.5.3.- Egypt

The Late Miocene record portrays a continuing withdrawal of the sea from the Egyptian domains, which culminated during the Messinian salinity crisis. In the Nile delta region, the Tortonian sequences reflect deposition in paralic and shelfal environments, associated with clinoform progradation in the eastern and subsidence in the western parts of the area (HARMS & WRAY, 1990). According to SAID (1990), the Late Miocene was a period characterised by erosion associated with the development of the oases and depressions of the Western Desert. Also the modern course of the Nile originated during the Late Miocene. The Gulf of Suez had lost its connection with the Mediterranean and marine ingressions into the Gulf of Suez - Red Sea rift system presumably arrived from the Indian Ocean. Sedimentation was characterised by the accumulation of thick evaporite sequences of the Zeit and South Garib (Gulf of Suez) and Abu Dabbab (northern Red Sea) formations.

II.5.4.- Tunisia

Two major orogenic phases of compressional tectonic deformation controlled the structural evolution from the Middle Miocene to the Late Pliocene. The oldest was late Middle to Late Miocene in age and induced thrusting in the north-western and folding in the eastern and central parts of Tunisia (BEN AYED, 1986; ROUVIER, 1977). The second main deformation phase ranged from the Late Miocene to the Late Pliocene and was characterised by folding of the orogenic nappe pile, relief steepening and concomitant molasse deposition. Increasing tectonic activity during the Middle and Late Miocene resulted in uplift and erosion in central Tunisia (mirrored by the coarse, fluviatile sands of the Beglia Formation; BUROLLET, 1956), reactivation of Tellian nappes (which were covered by shallow-marine and fine-grained deltaic sequences of the Saouaf Formation and its time-equivalents; BEN SALEM *et al.*, 1992) and reactivation of pre-existing (Cretaceous) fault systems, as well as in the development of NW-SE trending fault systems in central and offshore Tunisia (TURKI, 1985; ZARGOUNI, 1985; BEN AYED, 1986; CHIHI, 1995). Concomitantly, NW-SE directed compressional stresses caused rejuvenation of pre-existing fold systems and inversion of basins. The later, second deformation phase was also associated with folding and strike-slip motions along major basement lineaments all over the Atlas domains of Tunisia (CHIHI, 1995). In the north-west, a major phase of basin inversion was initiated in latest Miocene (Messinian) time (ROUVIER, 1977; BEN FERJANI *et al.*, 1990). As a consequence of these tectonic events, most of Tunisia emerged and was subjected to erosion. From west to east, three palaeogeographic subdomains can be distinguished, corresponding respectively to a fluvio-lacustrine, a shallow-marine and a coastal-marine environment of deposition.

In the various depositional domains of Tunisia predominantly siliciclastic successions were deposited (Serravallian to Messinian Kechabta, Oued El Maleh, Hakima, Oued Bel Khedim, Segui, Beni Khiar, Melqart, Somaa, Saouaf, Beglia and Birsa formations; BUROLLET, 1951 and 1956; BIELY *et al.*, 1972; BISMUTH, 1984; JEDDI, 1993; BEDIR *et al.*, 1996). Generally, it has not been

possible to date these clastics precisely, because of the lack of age-diagnostic fossils. In north-western Tunisia, the continental Kechabta Formation comprises fine and coarse-grained, molasse-type deposits (ROUVIER, 1977). The Segui Formation developed in central and parts of south-eastern Tunisia. It consists of conglomerates, marls and clays deposited in a variety of continental environments. In the Gafsa region, the sediments are mainly composed of conglomerates, sands and siltstones rich in vertebrate remains (Upper Beglia Formation; BIELY *et al.*, 1972). Coastal to shallow-marine, predominantly fine-grained terrigenous-clastic successions, including deltaic sediments, accumulated in the east (Saouaf Formation; BEN SALEM *et al.*, 1992). Locally, in the north-easternmost Gulf of Hammamet, relatively deep-water sequences with limestone and sandstone intercalations were laid down. These Tortonian successions (Terravechia Formation) contain rich foraminiferal faunas including the index species *Neogloboquadrina acostaensis*. They are time-equivalents of the coastal and shallow-marine Upper Saouaf and the fluviatile to continental sands of the Somaa Formation, which were deposited in the southern interior parts of the gulf (TAYECH, 1984). The Somaa Formation developed also in the eastern Sahel region. Here, it displays fine-grained, reddish sandstones with shale, sandy limestone and gypsum interbeds. Approximately time-equivalent sequences were laid down in the Gulf of Gabes.

23.- PIACENZIAN / GELASIAN (3.4 - 1.8 Ma)

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I.- MAIN FEATURES

I.1.- Time-slice definition and biochronology

This map was originally meant to portray the environmental and sedimentary-tectonic features of the Early Piacenzian, generally considered to correspond to the early Late Pliocene. Recently, however (RIO *et al.*, 1994), the post-Early Pliocene interval has been subdivided into the Middle Pliocene Piacenzian (~3.4 - 2.6 Ma) and the Late Pliocene Gelasian (~2.6 - 1.8 Ma; Fig. 17.1). Due to the overall non-marine character of the younger Pliocene Peri-Tethyan sequences it is often not possible to subdivide the Middle/Late Pliocene, i.e. to discriminate between the Piacenzian and the Gelasian. Consequently, the resultant map includes features which are pertinent to the interval from about 3.4 to 1.8 Ma. It thus displays the Pliocene palaeogeographic configuration and depositional environments which occurred after the late Early Pliocene large-scale tectonic reorganisation. In terms of regional stages of the Central and Eastern Paratethys, respectively Romanian and Akchagylian features are included. As far as mammal datings are available, they pertain to the latest Ruscinian - Villanyian interval (MN zones 15 - 17).

I.2.- Structural setting and kinematics

In the Middle to Late Pliocene, large parts of the northern and southern platform realms were subjected to regional uplift and the sea had withdrawn from by far most of the Peri-Tethys domains proper. In the Eastern Paratethys new basin configurations and subsidence regimes developed in the Middle to Late Pliocene, subsequent to (late) Early Pliocene tectonics. This resulted in the further uplift of the Greater Caucasus domain and in the origin of the NNW-SSE trending Akchagylian basin. Major features in the circum-Mediterranean parts of the Tethyan domain included the increased subsidence of the Tyrrhenian and southern Aegean back-arc basins and increased rates of uplift of the adjacent mountain chains in the course of the Middle/Late Pliocene and Pleistocene. In fact, the Middle and Late Pliocene vertical movements initiated the development of the present-day overall palaeogeographical and palaeotopographical features along the African - Eurasian collision zone.

I.3.- Outlines of palaeogeography and palaeoenvironments

In most of the Peri-Tethyan domains sedimentation had become confined to the accumulation of fluvio-lacustrine, clastic sequences in isolated basins during the Middle and Late Pliocene. However, in the Eastern Paratethys the palaeogeography and environmental / depositional conditions changed considerably. Here, the roughly N-S trending, locally deep Akchagylian basin (which extended as far north as 50 degrees latitude) originated. The sediments deposited in the region of the Black Sea depression indicate the existence of ephemeral connections with the Mediterranean Sea, as evidenced by the occurrence of horizons with calcareous nannoplankton. Possibly, the marine incursions entered through a corridor located in the north-eastern Aegean region. The most important changes in environmental and depositional conditions on the southern Peri-Tethys platform relative to the Late Tortonian involved a change from evaporitic towards (marine) clastic sedimentation in the Gulf of Suez - Red Sea basin system, around the Miocene - Pliocene transition. Shallow-marine environments persisted all along the northern margins of the African continent. In the Iberian - Moroccan domain the Strait of Gibraltar, which connects the Atlantic Ocean and the Mediterranean Sea, originated at the beginning of the Pliocene, after the Betic and Rifian corridors were closed in late Late Miocene (Messinian) time. The closure of these marine corridors resulted in the Messinian salinity crisis of the Mediterranean. In the course of the Late Miocene and Pliocene, almost

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everywhere on the northern and southern Peri-Tethys platforms river systems obtained courses and transport directions of clastics as they are known today.

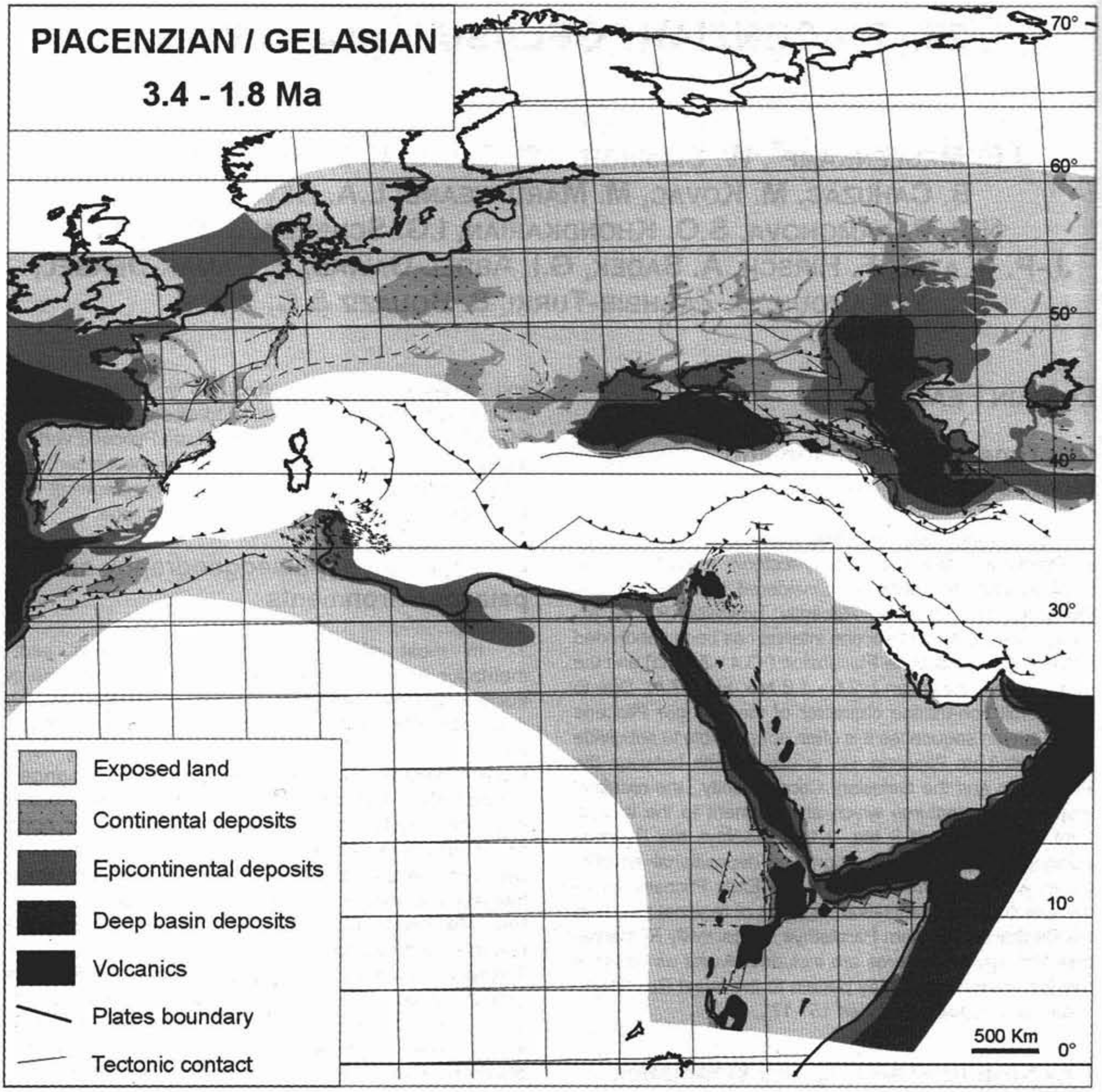

Fig. 23.1: Simplified palaeogeographic map of the Peri-Tethyan area during the Piacenzian / Gelasian.

II.- Description of Domains

II.1.- South-Western Europe

II.1.1.- Iberian Peninsula

All Iberian domains had emerged in the Middle and Late Pliocene, except for some peripheral basins. In these basins, located in the south-eastern part of Iberia (Almeria - Nijar, Carboneras and Murcia basins; Aguirre, 1998) and along the Atlantic coast, respectively littoral to shallow-marine platform sediments and interfingering alluvial and shallow-marine deposits accumulated. Continental strata were deposited in relatively small, intra-montane basins, such as those of Calatayud - Daroca and Teruel, Jucar - Cabriel, and Guadix - Baza (Agusti, 1986; Calvo *et al.*, 1993). In these basins sedimentation was characterised by shallow lake carbonates and fluvio-lacustrine clastics displaying well-developed pedogenic features. Sedimentation in the intra-montane basins was associated with radial extension, which co-occurred with localised volcanism and the overall uplift of Iberia. The regional uplift may have been related to approximately N-S compression, which supposedly not only affected the Iberian and Betic ranges (Sanz de Galdeano, 1996), but also the central parts of Iberia (De Vicente *et al.*, 1996). The geochronology of the Iberian Middle to Late Pliocene intra-montane continental sequences is based on the

superposition of micromammal localities, assigned to the MN 16 and MN 17 mammal zones. Additional geochronological evidence was obtained from the volcanic series emplaced in both the central and south-eastern parts of Iberia during the Middle to Late Pliocene (BELLON *et al.*, 1983).

II.1.2.- Aquitaine basin

Middle to Late Pliocene, marine sedimentation was confined to the westernmost part (i.e. present-day coastal area) of the Aquitaine basin and the south-eastern Bay of Biscay. Only neritic sands and marls were deposited, which mirror the final stage of westward progradation of shelf sediments. Locally, fluvio-lacustrine successions with lignite seams accumulated in the central, continental part of the basin (DUBREUILH *et al.*, 1995). Sedimentation of coarse(r)-grained clastics occurred near the southern margin of the basin and in the adjacent foothill areas of the northern Pyrenees. In the central Bay of Biscay pelagic oozes were deposited at bathyal depths (DSDP, leg 48; MONTADERT & ROBERTS, 1979).

II.2.- Western Europe

II.2.1.- European rift system

All along the Rhine graben tectonic activity strongly increased during the Early Pliocene. Coeval intra-graben deposits are absent as a consequence of uplift and concomitant erosion / non-deposition. Overlying the corresponding unconformity, deposition in fluvial and limnic facies was resumed during the Middle Pliocene in response to a new phase of rifting of the Rhine graben. In the south, conglomerates were deposited by the cmplex, rapidly changing Palaeo-Aare-Doubs river system, which extended from the Alps to the Bresse graben (VILLINGER, 1998). At the same time, the principal features of the present-day Rhine river system were established. In the Rhone graben, a similar Early Pliocene phase of tectonic uplift and erosion / non-deposition was also followed by renewed rifting during the later part of the Pliocene (RAT, 1984). The Bresse graben was transformed into a fluvio-lacustrine complex with swamps, in which also deltaic sands and conglomerates, as well as lignite-bearing marls, accumulated (SÉNAC, 1981; BONVALOT *et al.*, 1984). The coarser-grained clastics mostly originated from the Alps. They were mainly laid down by the Palaeo-Aare-Doubs and Rhone rivers. Fluvial conglomerates occurring along the western margin of the Bresse graben were derived from the adjacent Massif Central. Marine and brackish environments of deposition were completely absent in this part of the Rhone graben. However, marine clastics were deposited in a S-N orientated ria system that stretched all the way from the Gulf of Lions, until a short distance south of Lyon (BALLESIO,1972). This ria system developed in the incised Rhone valley in response to a marine transgression, related to the opening of the Gibraltar Strait at the Miocene - Pliocene transition and the subsequent flooding of the previously desiccated Mediterranean basin. Post-Miocene uplifting of highs, continuing development of volcanic build-ups and progressing erosive effects of rivers further enhanced the palaeomorphology of the Massif Central. The latter's pre-existing river system with northward-directed discharges of clastics was maintained. North of the massif, it included the Pre-Seine and the Cher as the most important rivers draining the massif.

II.3.- Central Europe / Carpatho-Pannonian region

II.3.1.- General features

In the Middle to Late Pliocene, major folding and thrusting of Outer Carpathian units related to subduction roll-back had come to a close. Only in the south-east, i.e. in the south-eastern parts of the Eastern Carpathians and in the adjacent area of the Southern Carpathians, foredeep sedimentation continued. Here, the Pliocene (Dacian - Romanian) sequences reach thicknesses of thousands of metres. They probably mirror high subsidence rates in response to increased and localised slab-pull. The present-day deep-seated seismic activity observed in the Vrancea area represents the expression of the final stage of subduction of the European plate.

In the intra-Carpathian domains of the Central Paratethys, Middle to Late Pliocene sedimentation was mainly confined to (local) accumulation of continental clastics. However, some depocentres in the south-western part of the back-arc basin system were filled with Pliocene deposits reaching ticknesses ranging from some hundreds to about 1000 metres (Drava and Sava basins; MEULENKAMP *et al.*, 1996). Subsequent to the Late Miocene, alkali-basaltic volcanism played an important role in the intra-arc domains during the Plio-Pleistocene. This volcanism was widespread and occurred in the Styrian and Danube basins, Transdanubian Central Range area, Great Hungarian Plain, South Slovakian - North Hungarian volcanic domain, and south-eastern Transylvania (SZABO *et al.*, 1992).

II.3.2.- Regional aspects

The Middle / Late Pliocene evolution of the Carpatho-Pannonian domain was characterised by orogenic uplift and tectonic inversion of the intra-Carpathian regions. This was followed by the development of river systems similar to those of the Present and of "residual lakes". In the northern and central parts of the back-arc basin predominantly fluviatile and fluvio-lacustrine sequences were deposited. They are composed of gravels, sands, silts and clays; their thicknesses range up to several hundreds of metres. The Middle/Late Pliocene stress field was similar to that of the present-day (BADA, 1999); it was characterised by compression perpendicular to the orogenic trend. Within the Western (and probably also Eastern) Carpathians, KOVAC *et al.* (in press) noted an areal change from compression in the uplifted Outer Carpathian belt to arc-parallel extension in the intra-Carpathian region.

In the Dacic basin (which belonged to the domain of the Eastern Paratethys), the interval mapped corresponds to the (Middle) Romanian which is correlative to the Piacenzian (PAPAIANOPOL & MARINESCU, 1995). The fill of the Dacic basin shows a significant change in faunal composition (extinction of Limnocardiids and appearance

of faunas rich in Unionids, Viviparids, and Melanopsids), which portrays the further decrease of surface-water salinities during the Pliocene. The Dacic basin developed "on top of» the external zone of the Outer Carpathians and the northern part of the Moesian platform. Its fluvio-lacustrine sediment successions are fairly similar to those deposited farther to the west, i.e. in the south-eastern part of the Pannonian back-arc realm. They consist of conglomerates, gravels, sandstones and clays. Coal seams are locally intercalated; their accumulation was far more important earlier, i.e. in Dacian to Early Romanian times. Faunal differences indicate that the depositional domains of the Pannonian intra-arc region and the outer-arc Dacic basin were completely separated by the Carpathian chain ("Upper *Paludina* beds" and "sculptured Melanopsid beds", respectively). In the easternmost parts of the overall fluvio-lacustrine Dacic basin, evidence is found of ephemeral marine ingressions, from the adjacent Euxinian/Black Sea domains in the east. Such ingressions are indicated by the occurrence of brackish-water molluscs (*Euxinicardium, Adacna, Didacna* and *Monodacna*; PAPAIANOPOL *et al.*, 1995) and calcareous nannoplankton associations indicative of zones NN15 - NN16 (MARUNTEANU & PAPAIANOPOL, 1995).

II.4.- Eastern Europe / Western Asia

II.4.1.- General features

Since latest Miocene (Late Pontian) time, the Eastern Paratethys consisted of two major basins, the Dacic - Euxinian basin system and the Caspian basin. The Pliocene faunas of the Dacic - Black Sea basin system were inherited from the Pontian. The brackish Kujalnician basin corresponded approximately with the Black Sea area and the Gulfs of Azov - Kuban and Riony. In the Early Pliocene, the Caspian basin experienced a major regression, which was accompanied by a basin-wide reduction of the salinity. The pre-Akchagylian Balakhanian basin included the South Caspian depression and the Kura Gulf. Sea level lowering resulted in deeply incised valleys (up to a few hundred metres) of the Palaeo-Volga, Kura and Amudarja rivers. The thus rugged relief was invaded by the Middle to Late Pliocene Akchagylian Sea; the origin of this transgression is unknown. The Akchagylian Sea was characterised by reduced salinities and euryhaline biotas. Numerous new endemic taxa of molluscs (PARAMONOVA, 1994), ostracods and diatoms evolved from marine (Mediterranean) ancestors. Pollen assemblages and leaf remains indicate the occurrence of a forested hinterland with conditions similar to those of the present-day taiga. They also suggest an environmental change from a cool and dry climate towards one typified by relatively warm and wet conditions ("broad-leaved forest zone"; NEOGENE SYSTEM, 1986).

II.4.2.- Regional aspects

II.4.2.1.- East European platform and Scythian plate

The major part of the western East European platform corresponded to the Russian Highland, which was intersected by branched river systems, distributing sandy sediments during the Akchagylian (Palaeo-Don alluvial deposits; Nagavskaya and Krivskaya Suites yielding Middle/Late Pliocene micromammals). Tectonics caused a lowering of the Precaspian depression and concomitant deep incision of the bordering rivers, which were subsequently affected by Middle and Late Akchagylian marine transgressions. Relatively deep-water, clayey deposits (350 - 500 m) accumulated in the central parts of the Precaspian depression. Northwards, the Akchagylian Sea ingressed the Volga and Kama basins. The poor and uniform, euryhaline faunal assemblages of these northerly basins reflect low salinities as compared to those of the main basin (JAKHIMOVITCH *et al.*, 1985; NEOGENE SYSTEM, 1986).

II.4.2.2.- Turanian plate, Tien Shan, Pamirs and Kopet Dagh

The slopes of the Ural and Mugodzary displayed incised river valleys. The terrestrial high of the western Turanian plate, southern Mangyshlak and Ustjurt plateau was bordered to the south by the deep-water Fore-Kopet Dagh depression. This depression was filled with terrigenous-clastic, mainly marine sediments (up to 1000 m). In the relatively shallow parts of the western Kopet Dagh Gulf, calcareous sediments with rich endemic faunas (*Avicardium, Miricardium, Andrussella, Avimactra)* accumulated in Middle Akchagylian time. Marine terrigenous-clastic and evaporitic sediments accumulated contemporaneously in the South Aral Gulf, which was probably connected with the Kopet Dagh Gulf by a narrow corridor. During the Middle Akchagylian, marine ingressions reached as far as the Palaeo-Murgab, the Tedzhen and Amudarja valleys. Here, the Middle Akchagylian sequences are succeeded upwards by deltaic and alluvial deposits (Repetek Suite). The coarsest molasse-type clastics developed around the eastern Kopet Dagh, Pamirs and Tien Shan orogenic belts (Polisak Suite of the Tadjic and Fergana depressions, east of mapped area). The northern Karakum area was a highland where thick eolian successions were deposited. These successions contain sandy material derived from the older Neogene. The Palaeo-Zeravshan and -Sarysu rivers transported clastics from the southern Kazakhstan highland to the south. These clastics eventually accumulated in the lacustrine Aryskum depression and along the northern margin of the Karakum. Sandy to clayey lake sediments were also deposited in the Teniz depression. Fluvio-lacustrine, silty and clayey sediments were deposited in the Turgaj area and in western Siberia (Kustanay Suite, Bitekey beds; NEOGENE SYSTEM, 1986).

II.4.2.3.- Black Sea depression and Caucasus

In Akchagylian time, the brackish Kujalnician Sea (part of the Black Sea depression) had no direct connection with the world oceans, but there was a corridor connecting it with the Akchagylian basin. Shallow-water terrigenous clastics with Kujalnician faunas are known from the Riony depression. A new episode of orogenic movements in the Greater Caucasus resulted in a pronounced increase of clastic supply from the evolving chain towards the bordering plains. Seismic data portray the growth of prograding fans into the Terek - Caspian

and eastern Kubanian molasse foredeeps (KUNIN *et al.*, 1989). Thick successions of coarse terrigenous clastics were laid down in the Kura intra-mountain depression (up to 900 metres of continental conglomerates, sands and aleurites of the Alazan Series). To the east these deposits pass into shallow-marine sediments, partly resembling "contourites". Deep-water muds (50-100 m) accumulated in the Kobystan part of the South Caspian depression. Eruptions in the Kazbek area of the Central Caucasian and Adzharo - Trialet ridge produced abundant volcaniclastics, such as, for instance, those of the Ruch-Dzoar Suite in the Palaeo-Terek delta (NEOGENE SYSTEM, 1986).

II.5.- Southern Peri-Tethys platform

II.5.1.- Saudi-Oman domains of the Arabian peninsula

The distribution of Plio-Quaternary deposits on the Arabian peninsula was mainly controlled by palaeotopographical (and climatic) conditions. Erosion products accumulated in a continental basin occupying the central regions (Rub' al Khali and interior Oman); the local thicknesses of these sedimentary sequences are proportional to the elevation of the surrounding relief. The age of the oldest alluvial fan deposits is not well constrained; most likely, they are of Pliocene age (CHEVREL *et al.*, 1992). The youngest channel conglomerates south of the Oman mountain chain (Barzaman Formation) are also tentatively assigned to the Pliocene. These deposits are correlative to conglomerates deposited near the shore by wadis draining the coastal regions. They have their widest distribution in the coastal plains of Batinah (in the north), Batain (in the north-east), and Salalah (in the south).

II.5.2.- Israel

The Late(st) Messinian sea level drop resulted in the origin of a submarine canyon along the Levantine coast (DRUCKMAN *et al.*, 1995). Subsequently, open-marine marls of the Yaffo Formation were deposited in the coastal plain area in response to the Early Pliocene flooding. However, the sea did not re-invade the Dead Sea Rift basin. The Middle to Late Pliocene evolution of the Levantine domain was characterised by the development of a large fluvio-lacustrine system, which discharged in the Dead Sea Rift basin. Lakes occurred in the regions of the Negev/Sinai and Arava.

II.5.3.- Egypt

In response to the Early Pliocene flooding the sea invaded the northern coastal areas, including the northern Nile delta region, and the canyon of the Eonile, as far south as Aswan. This transgression was followed by regression in Middle/Late Pliocene times. Fluvio-marine sediments (Wastani Formation) were deposited in the Nile delta region and in Wadi El-Natrun (Gar El-Muluk Formation); fluviatile sediments (Muneiha and Issawia formations) were deposited along the Nile valley. The Pliocene in the coastal areas of the Mediterranean west of the delta is represented by pink limestones of shallow-water origin. In the Cairo - Suez area white, porcellaneous limestones with flint (Hamzi Formation; SAID, 1971) accumulated in fresh-water, lagoonal environments during a humid episode in the Late Pliocene. This major climatic change is evidenced by fluvial drainage patterns indicative of relatively humid conditions (SAID, 1990). In the Pliocene, the Gulf of Suez and Red Sea were connected with the Indian Ocean via the Strait of Bab El-Mandab (SAID, 1990; ISSAWI *et al.*, 1999). In the Gulf of Suez fossiliferous, predominantly clastic, sequences were deposited during the Late Pliocene; up to about 1000 m of sands and anhydrite were encountered in wells drilled in the gulf. Late Pliocene, marine calcareous sandstones and marls (Shagra Formation) yielding Indo-Pacific faunal associations, were deposited along the margins of the northern Red Sea.

II.5.4.- Tunisia

The Middle and Late Pliocene is especially well-documented in the coastal region of Nabeul - Hammamet (north-eastern Tunisia). In this area it consists of sands, sandy marls and clays (including the "Argiles de Sidi Barka"), which are overlain by the partly turbiditic "Sables et grès de Hammamet". The succession contains rich assemblages of benthic and planktonic foraminifera (including the index taxa *Sphaeroidinellopsis subdehiscens*, *Globorotalia puncticulata*, *G. inflata* and *Globigerinoides obliquus extremus*), which allow a correlation with the MPL4 - MPL6 Mediterranean planktonic foraminiferal zonal interval (COLLEUIL, 1976; BESÈME, 1979; BISMUTH, 1984; DAMAK-DERBEL *et al.*, 1991; BEN SALEM, 1992). Five palaeogeographical / depositional domains can be recognised covering the region located between the emerged Tunisian part of the African continent and the Pelagian Sea. These include the exposed land, as well as fluvio-lacustrine, shallow-marine, coastal-marine and deep-marine depositional environments. One domain comprises the Khroumirie mountains (northern Tunisia; BUROLLET, 1951), the central and southern Atlas mountains, and eastern and south-eastern Tunisia (Sahel and Jeffara coastal plains). In areas adjacent to this domain, fluvio-lacustrine deposits accumulated. They consist of clays, sandy clays and conglomerates. Locally, these deposits include beds of gypsum and limestone. These sequences were deposited in environments often resembling those of the present-day coastal plains of eastern Tunisia. In the Sahel region fluvial deposits reach a thickness of 800 metres. The fluvio-lacustrine depositional environments passed into shallow-water, partly lagoonal environments with fluctuating salinities. In the latter, shales, sandstones and limestones accumulated in association with evaporitic interbeds (gypsum, anhydrite) and calcareous sediments yielding bryozoa, echinoids, pelecypods and benthic foraminifera.

Altogether, these coastal to shallow-marine environments of deposition were situated in a SSE-NNW trending zone extending from the Gulf of Gabes to the Gulf of Tunis, including north-eastern Tunisia (area between Tunis and Bizerte, and Cap Bon; HOOYBERGHS, 1977) and eastern Tunisia (area between Sousse and Mahdia; KAMOUN, 1981). In the Gulf of Tunis, predominantly sandy, fossiliferous sediments accumulated; more than 800 m were recorded in the Raoued 1 well. In the Gulf of

Gabes, some hundreds of metres of mainly bioclastic limestones and sandy to silty clastics and, locally, gypsiferous clays were deposited. The circa-littoral, Middle to Late Pliocene sequences of the Gulf of Hammamet consist of sandy and glauconitic, fossiliferous clays (containing the index species *Globorotalia crassaformis* and *G. inflata*), bioclastic limestones and lumachelles. Here, the NW-SE trending Jriba depocentre contains 800 m of Middle/Late Pliocene sediments. Farther offshore, in the domain extending from the Gulf of Hammamet to the Gulf of Tunis, deep(er)-marine clays and silty marls were deposited. The central part of the Gulf of Hammamet corresponds to another, SW-NE trending depocentre. The Middle/Late Pliocene tectonic framework of the Cap Bon and Sahel regions was characterised by a SW-NE trending extensional stress regime, as evidenced by N 140-oriented syn-sedimentary conjugate normal fault systems (COLLEUIL, 1976; KAMOUN, 1981; DAMAK-DERBEL, 1993).

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24.- LAST GLACIAL MAXIMUM (LATE PLEISTOCENE, 22000 - 18000 YEARS BP)

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WITH COLLABORATION OF
R. RAGALA & B. VRIELYNCK

I.- MAIN FEATURES

Changes since the previous map

The Last Glacial Maximum (LGM) map not having a counterpart in the Tethys atlas (DERCOURT *et al.*, 1993), the Tethyan and Peri-Tethyan regions have been considered as a whole and mapped together. This map features one of the best-documented palaeoclimatic periods of the Quaternary whose study gives clues about past, as well as expected forthcoming evolutions of the environment (BRULHET, PETIT-MAIRE *et al.,* 1999). However, it cannot be considered as a conclusion for the Atlas since it does not represent an end-point – the same could be said for the Early Holocene or the current situation – but rather one of the numerous paroxysmal events having characterised the last millions of years in the Earth's evolution. This map can be better defined as an illustration of the strong palaeoclimatic influences that appear with respect to other geodynamic controls when the studied time slices become very short.

Though active tectonics and volcanism characterise many areas of the Tethyan domain or even the Peri-Tethyan platforms, present geological and morphostructural patterns can be roughly considered as unchanged during this period, except for the rapid isostatic effects of deglaciation under and around the main LGM ice sheets. This is why the geological framework represented here is the current one.

About 2 Ma separate the Piacenzian map, the last Tertiary map of the Peri-Tethys Atlas, from this map that takes place around 20 000 years ago. During this period no major geodynamical change occurred in the remnant Peri-Tethyan domain where the African - Arabian and Eurasian plates confront. The major Cainozoic orogenies related to the convergence between these plates were either almost achieved before Late Pliocene, or already started before this time and are still active. In Western Europe, the Alpine and Pyrenean orogenies ended in Early Pliocene and Late Oligocene, respectively. The main tectonic phases of the western Mediterranean chains (Maghrebides, Betic Cordillieras), are mainly Miocene in age, whereas in Central Europe, the Carpathian fold and thrust belt is essentially Miocene. In the Balkan area, the Hellenides orogen is achieved in Oligocene and the Balkanides and Rhodope tectonics since the end of Eocene. These inactive Cainozoic orogenic belts induced by the convergence of the African/Apulian/Arabian plates and the Eurasian plate resulted in the uplift of large areas of the northern and southern Tethyan margins. Since the end of the major tectonic phases, these orogens are mainly submitted to isostatic adjustments and to erosion.

The major active zones of Late Pliocene are very similar to the present day deformation zones. Even if the geometry of these active tectonic features slightly changed since 2 Ma, the geodynamical context remained stable during this period. In the eastern domain, the active orogenic belts (Crimea - Great Caucasus, Taurides, Pontides, Lesser Caucasus, Zagros) resulting from the Arabia - Eurasia collision were active prior to the Pliocene. The Aegean and Calabrian subductions, initiated in Early Miocene, probably together with the associated extensions of the Aegean and Thyrrenian domains. The major active strike-slip faults of the eastern Mediterranean area, the right lateral North Anatolian fault and the left lateral Levant fault, related to the collision between Africa - Arabia and Eurasia, already existed in Late Pliocene with the same kinematics, even if their detailed geometry and displacement rate slightly differed at that time. So, within the two last million years, only minor environmental change of tectonic origin can be expected.

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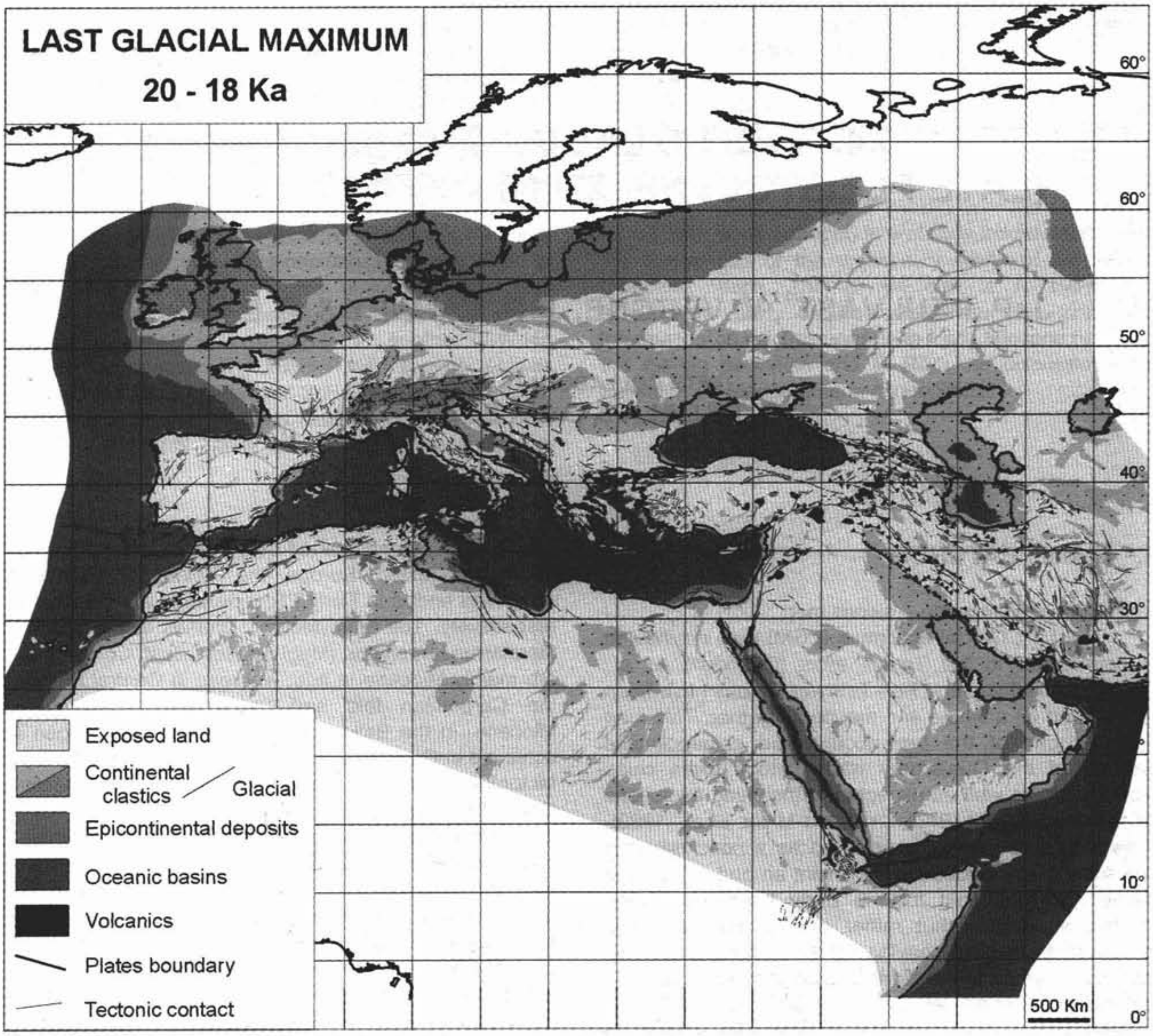

Fig. 24.1: Simplified map of the Peri-Tethyan area during the LMG.

Global climatic changes are the main factors of the spectacular differences between the LGM situation and the Piacenzian or the Present environmental patterns (IMBRIE, 1985; EHLERS, 1996; WILSON *et al.*, 1999). The LGM and subsequent periods form only the last sequence of a long series of sharp fluctuations that characterise the Late Pliocene and the Quaternary.

The comparison of long terrestrial sequences with deep ice cores and deep sea sediment cores – which give more continuous stratigraphic records –, in particular by the means of oxygen-isotope analysis, reveals that about eight glacial-interglacial cycles occurred during the past 0.8 Ma, each cycle lasting ca. 100 000 years (according to the Milankovitch theory; SHACKLETON, 1987; BERGER, 1988; MOMMERSTEEG *et al.*, 1995; TZEDAKIS *et al.*, 1997). Complemented by data from deep ice cores (Antarctic, Vostok: LORIUS *et al.*, 1988; JOUZEL *et al.*, 1993 and 1994; Greenland, GRIP, GISP 2: Greenland ice-core project Members, 1993; DANSGAARD *et al.*, 1993; GROOTES *et al.*, 1993; PETIT *et al.*, 1999; WILSON *et al.*, 1999), this comparison shows that the last pleniglacial period reached a peak between 25 and 18 ka B.P. Compared to the present one, the main characteristics of this period are:

1.- a negative thermal budget, in response to the high-latitude decrease of solar insolation, and a low concentration of carbon atmospheric reservoir (pCO_2 = 200 ppm, 75% of the pre-industrial value, or 50% of Present);

2.- the rapid and drastic accretion of ice in extent and volume (nearly 30% of the surface of the continents covered by ice: COHMAP, 1988; Poland: very rapid, surge-like advance, KOSARSKY, 1995);

3.- the alteration of the hydrological cycle by changes and transfers in water (ice - ocean - air) reservoirs (STARKEL, 1995);

4.- as a consequence, latitudinal displacement of climate - environment system: pressure zones, precipitation belts, soil groups, vegetation regions, etc. (MCINTYRE, KIPP *et al.*, 1976, FRENZEL *et al.*, 1992; PETIT-MAIRE *et al.*, 1994); southern translation of robust atmospheric-oceanic eastward forcing; lowering of sea-

level (-100/130 m below present surface, next to the shelf break; SHACKLETON, 1987); spectacular changes in hydrography (e.g., fluvial main divides, problematic marine and lacustrine interbasinal connections), and physiography (e.g., continental ablation, water and/or eolian discharges to the ocean, preponderance of lithic sedimentation); a widespread retreat of human settlements from the coldest areas, etc.

The time slice: a discussion

The LGM is part of the Late Quaternary, a time-interval corresponding approximately to the last 130 ka (Interglacial – Glacial). The Quaternary period is subdivided into sections which are usually defined by the Matuyama - Brunhes geomagnetic polarity reversal (limit Early - Middle Quaternary: 790 ka B.P.) and the beginning of oxygen isotope substage 5e - the last Interglacial, at the limit Middle - Late Quaternary: 130 ka B.P. The complex glacial sequence (stages 5d to 2) that occurred during the time interval between this last limit and 10 ka B.P. (end of Last Glacial) has received glaciation names assigned to geographical domains: Weichselian in NW Europe, Devensian in the British Isles, Würm in Central Europe, Valdai in European Russia. It is usually subdivided into three stages (Early, Middle, Late) linked to the marine isotope chronology (SHACKLETON, 1987; SHACKLETON *et al.*, 1990). According to current knowledge, the last part of the Late Weichselian glaciation started around 28 ka B.P..

The LGM dates are generally considered to lie between 22 and 17 ka B.P. (stage 2; BRULHET, PETIT-MAIRE *et al.,* 1999) and has been recently defined between Heinrich event 2 and Heinrich event 3. Nevertheless, FRENZEL *et al.*, (1992) stressed the fact that the most appropriate climate for the expansion of glaciers and ice sheets prevailed some time before the maximal extent of the inland ice masses occurred, and that once the inland ice had reached this extension the climate seems to have already favoured its disintegration. Moreover, dating methods and resolution vary with each author and more or less important lags occur in regional – or even local – reactions to global changes of temperatures. The map therefore represents the effects of the most intense climatic signal rather than a single moment of development. The time-slice may appear very short (4000 to 5000 years) but it cannot be as precise as suggested by a notion of Maximum, which, in fact, is time-transgressive. The problem is different for the tectonic and magmatic (volcanic) events. For them, such an interval has no precise meaning and the features represented here correspond to a situation which prevailed during the whole Late Pleistocene and lasted until the Holocene and the Present.

Recent works on the Late Pleistocene, in particular studies on deep ice and deep sea cores, reveal many short-term fluctuations which have made the definition of the time slice represented on the LGM map a rather delicate process. The plot of $\delta^{18}O$ reveals 24 interstadials between 115 ka and 14 ka (stages 5d to 1). The repeated episodes of rapid warming/cooling are known as "Dansgaard - Oeschger oscillations" (rapid increases in temperatures over ice sheet during ca 100 years follow by relatively slower cooling). These shorter events are bundled together into longer cooling cycles - Bond cycles - characterised by a steady drop between successive peaks in the $\delta^{18}O$ values. The large ice rafted debris peaks (IRD) at the end of each Bond cycle are "Heinrich events". Heinrich events were first recognised as 'ice surges' into the North Atlantic, caused by sudden collapses of the major ice sheets. They occurred around ≃ 55 ka (H_2), ≃ 44 ka, 35.2 ka, 27.2 ka, 22.2 ka and 15 ka B.P. Each of them lasted *ca* 2000-2500 years, with both Heinrich events 2 and 3 representing most probably the *extreme* LGM conditions mapped here. If this is the case, slightly milder (though still more arid than present) conditions may have prevailed during parts of this period (BOND *et al.*, 1992 and 1993; DANSGAARD *et al.*, 1993; BOND & LOTTI, 1995; HUNTLEY *et al.*, 1995; WILSON *et al.,* 1999; GROUSSET *et al.*, 2000 and e.g., GUIOT *et al.*, 1993; TURON *et al.,* 2000).

The Peri-Tethys LGM map differs from recently published palaeoenvironmental maps of the Quaternary because of its predominant geological component. According to the common Peri-Tethys legend, the palaeoenvironments are mainly depositional, with corresponding lithological facies. Some palaeoclimatic data are represented, but not vegetation zones (FRENZEL *et al*, 1992; PRENTICE *et al.,* 1993; BRULHET, PETIT-MAIRE *et al.*, 1999; PETIT-MAIRE & BOUYSSE, 1999). On the contrary, the tectonic and volcanic context appears as a major geodynamic component – as it actually implies longer time slices –, whereas this is not the case for other Quaternary maps.

Deposits were mapped from existing geological maps, e.g., the UNESCO Quaternary maps of Europe (and Maghreb) at 1:2 500 000 and more general maps in other regions (see VELICHKO & SPASSKAYA, 1991; the Geological Atlas of the World at 1:10 000 000). More precise thematic, regional or local publications were used when available (references in the text), though the scale of the map did not allow the representation of all details (e.g., it was impossible to draw alluvial deposits along many valleys such as the Seine, the Loire or the Meuse rivers). In these publications, many deposits are dated as Late Pleistocene, if not Pleistocene, especially alluvial, eolian or marine deposits that have been reported here though they cannot be considered as strictly related to the LGM. Their geographical extent is assumed to be almost the same as the presently visible one in spite of locally important erosion (e.g., in the loesses and sands of Northern Africa: COUDÉ-GAUSSEN & ROGNON, 1986). Most limits, such as the coastlines, the glacier or permafrost, as well as climatic features, are more clearly related to the LGM - though some of them are time transgressive, i.e. glacial limits. Some domains, however, could not be precisely outlined on the basis of geological limits as in the case of arid regions, where eolian activity occurred continuously and repeatedly during the whole Quaternary, and where all related deposits were represented, even though parts of

them might have been covered by steppic vegetation during the LGM (parts of North-Western Africa: ROGNON, 1990; TURON *et al.*, 2000).

II.- STRUCTURAL SETTING AND KINEMATICS

Among the 24 maps of the Peri-Tethyan Atlas the LGM map is a particular one. Its time-slice of 4 000 years is very small (about 3 orders of magnitude) compared to the mean time-slices of 2 Ma of the Mesozoic and Tertiary maps. This point is fundamental as far as the tectonic aspect of the LGM maps is concerned. Generally, the age of a tectonic event (beginning and end) can reasonably be dated with an accuracy of several millions of years. This accuracy depends from the type of tectonic event (extensional, compressional), the regional stratigraphy, and the available data (field and/or subsurface data). As one single tectonic event commonly lasts several millions of years for the Mesozoic and Tertiary maps the duration of the time-slices generally allows to integrate one tectonic event (or a lack of tectonics) in one region for a given map. If we consider the two last maps of the Peri-Tethys Atlas, the Piacenzian and the LGM maps, they are separated by a period of about 2 Ma almost equal to the mean time-slice of the PTP maps. Because no major change in the tectonic context occurred during the last 2 Ma, the Piacenzian and the LGM maps should be almost identical in the approach adopted for the other Peri-Tethys maps. Thus, it appears that a different tectonic approach has to be used for the LGM map that requests (1) a better accuracy in determining the age of the tectonic patterns, and (2) a more precise geometry of the zones of deformations (major faults, subduction zones, fold and thrust belts).

Determining the real deformation pattern during the period 22 000-18 000 years B.P. is an unrealistic objective. First, a large part of the sediments deposited during this time-slice has been submerged since the end of the last glacial period, and second the formations accurately dated from this period are too scarce to obtain a sufficient covering of the map. An attractive possibility was to consider the LGM map as a seismotectonic map. Such a seismotectonic approach allows to elaborate precise detailed maps of the active tectonic features using the surface ruptures and the instrumental and historical seismological data. Nevertheless the seismotectonic approach integrates a very short period of time restricted to several centuries and many faults active in the Late Pleistocene were not necessarily active during the historical period. So, rather than limiting the tectonic aspect of the LGM map either to the exact time-slice of the map or to the historical period, we integrated the tectonic deformations since the Late Quaternary (about the last 130 000 years). Using this time-slice, we consider the tectonic pattern during the last interglacial-glacial period. Available tectonic data about this well studied period are abundant and complement the seismotectonic data. The result may be defined as a neotectonic map of Late Quaternary. Because of the quality of the original data, the level of detail of the tectonic features reported on this map is appreciably better on this map than on the other PTP maps and in accordance with the palaeoenvironmental map.

Different categories of geodynamical and tectonic data have been retained:

- several major active plate boundaries are known from marine data. In the Mediterranean Sea, the Calabrian and the Aegean subduction zones including the associated accretionary wedges (Calabrian prism and East Mediterranean ridge) have been incorporated in their present configurations as well as the accretionary ridges of the Rea Sea and of the Gulf of Aden;
- the major active strike-slip fault systems related to the Eurasia - Arabia collision, the right lateral North Anatolian fault and the left lateral Levant fault, have been mapped from the abundant literature;
- the active tectonic features (faults and folds) of the complex Arabia - Eurasia collision zone, including the East Anatolian fault zone, the Crimea - Caucasus area, the Lesser Caucasus and Alborz fold belts, the Apsheron ridge, the Zagros fold and thrust belt, and the active shear zones (essentially strike-slip faults) bounding the Iranian blocks, are mainly drawn from local data;
- the tectonic setting of the well documented active zone surrounding the western Mediterranean Sea in Maghreb, Iberian Peninsula, and Apulia, mainly composed of reactivated Cainozoic structures, is synthesised from the regional seismotectonic and neotectonic maps of Morocco, Tunisia, Maghreb, Italy and of the Mediterranean domains, and from seismotectonic and neotectonic works as well;
- the diffuse pattern of active and recent faults in the Western European platform (France, British Islands, Germany) is reported from the data provided by the national geological surveys of these countries. Note that the seismic active faults related to the isostatic effects of deglaciation, especially in Great Britain and Scandinavia are not reported because these faults were obviously not active during the Last Glacial Maximum;
- the recent tectonic activity in the major Tertiary orogens (Alps, Pyrenees) results mainly of the reactivation of inherited Cainozoic faults, and of uplifts probably due to isostatic adjustments. In these regions, the data are issued from regional neotectonic studies. In the Dinarides the neotectonic activity is poorly documented. We only mentioned the major Cainozoic structures that can be associated with a seismic activity.

Our objectives were to present a tectonic setting that may (1) provide a clear view of the tectonic activity during Late Quaternary, and (2) complement the associated palaeoenvironmental map. So, many data that should be included in a pure neotectonic map do not exist here (kinematics, vertical movements, morphology, palaeostress fields, etc.).

III.- DEFINITION OF THE LGM DOMAINS

The definition of domains is mainly based upon climatic zonation (FRENZEL *et al.*, 1992). The geological and morphostructural subdivisions will be considered in more detail in the next section (description of domains).

III.1.- The palaeoclimatic zonation

The CLIMAP and SPECMAP reconstructions (1976, 1981; MCINTYRE, KIPP *et al.,* 1976), the later general circulation models (GCM), the PMIP (Palaeoclimatic Modeling Intercomparison Project) and the PAGES (Past Global Changes) etc. give a more and more precise picture of the climatic zonation in the Tethyan and Peri-Tethyan regions during the LGM (BERGER, 1996; THIEDE & BAUCH, 1999). According to these models, the global temperature was 4.5°C lower than today, and the areas south of the major ice sheets were relatively arid. In contrast, reduced evaporation in Central Asia may have given rise to a regional increase in available moisture.

On continental areas, the presence of large and thick ice sheets resulted in the development of individual permanent high-pressure cells. The clockwise circulation of the anticyclone surface airflow extended far around the Fenno-Scandian ice sheet and played a major role in the formation of extensive loess and windblown sand deposits in North-Western, Central and Eastern Europe (LAUTRIDOU, 1985; PÉCSI, 1992; ANTOINE *et al.,* 1998). The southward extension of this permanent anticyclone caused a southward shift of the mid-latitude cyclones, so that the main supply of precipitation to the ice sheets would have taken place along their southern margins; as a result, ice sheets could have likely advanced towards the south while retreating at higher latitudes. In continental areas, where moisture was insufficient for glacier development - the available moisture index and annual precipitation were - 60 ± 20 % north of Mediterranean sea. The enhanced negative heat budget at the ground surface resulted in the very cold conditions of the periglacial areas - the drop of annual mean air temperatures was about 8 to 15°C or more in Western Europe (FRENZEL *et al.*, 1992; PEYRON *et al*, 1998; JOUSSAUME & GUIOT, 1999) and in aggradation of permafrost along a 600 km wide zone in Europe.

Sea ice developed in the North Atlantic as far south as 40-45°N in winter (in summer, the southern edge of permanent pack ice was located near 60°N: CLIMAP, 1976, or even at higher latitudes, with some input of warm water resulting in sufficient air humidity for providing snow to the ice sheets: PETIT-MAIRE & BOUYSSE, 1999). The southward displacement of the oceanic polar front led to a corresponding displacement of mid latitude cyclones and to a marked reduction in the formation of deep water in this ocean, which in turn led to important changes in global climate (BROEKER & DENTON, 1990; WILSON *et al.,* 1999). The warming influence of the Atlantic Ocean on the climates of Europe appears to have been negligible (FRENZEL *et al.*, 1992).

III.2.- Main domains and environments

III.2.1.- Northern glaciated environments

From offshore Scotland and Northern Ireland ice cap to the Feroe Ridge, the coasts and ocean are exposed to the effects of diverging-anticyclonic winds and the probable permanent congelation of sea surface (pack ice). Worth to note is the almost submeridian limit of permanent sea ice.

Inland, two main ice sheets cover the British Isles and Fennoscandia (VELICHKO & FAUSTOVA, 1992). Their maximum extension, such as represented here, is time transgressive, from 24 to 17 ka B.P. It corresponds to the Dimlington stage (British Isles), to the Haugesund (Rogne) maximum advance (Norway), and to the Ne maximum advance (Denmark) and is represented in Northern Germany and Poland by the Brandenburg, Poznan and Pomeranian morainic belts. In the Russian plain of Eastern Europe, it corresponds to the Late Valdai, which is the coldest phase of the Valdai glaciation. Centred in the mountains of Scotland, Cumberland, Wales and Ireland, the Late Devensian glaciers did not cover South and South-West England, but extended beyond the present day shoreline onto the emerged shelf of the North Sea. The Scandinavian ice sheet covered the east of this area, beyond the deep Norwegian Trench, but, according to some authors, a broad zone in the Central North Sea seems to have remained ice free, whereas others maintain that a contact may have existed between the Scandinavian and British ice sheets between 29.4 and 22 ka B.P., i.e. before the Dimlington stage (SEJRUP *et al.*, 1994; LAMBECK, 1995a and b, 1996b). When the ice reached its maximum limit on the European continent, the confluence in the North Sea was broken.

III.2.2.- Mid latitude environments

III.2.2.1.- Periglacial environments: general

Several cold phases occurred during the Late Pleistocene outside the glaciated areas (PISSART, 1987). In central Russia, they are recorded in three horizons of ice wedge pseudomorph and other cryogenic structures among which the youngest (20-17 ka B.P.) is the best preserved (VELICHKO & NECHAEV, 1992). At this time, the permafrost reached its maximum expansion (VANDENBERGHE & PISSART, 1993; VAN VLIET-LANOË, 1998). The southern limits of the periglacial zone are associated with sea ice limits and appear to be partly controlled by topography within the continents. In European Russia, the permafrost spread 2000 km south of its present border (KONDRATJEVA *et al.*, 1993). Its thickness attained 200-400 m in the Moscow area. In Central and Western Europe, most of the areas north of the Garonne river, the Alps and the Carpathians were characterised by continuous permafrost, 10 to 100 m thick, with annual mean temperatures <-3 to -5°C. Well

known here from ice-wedge polygons, frost fissures or ice veins, its geographical extent is more uncertain in southern Europe and in Central Asia. In all cases, the limits remain a subject of discussion and the outlines represented here, more southern than the limits chosen by FRENZEL *et al.* (1992), may substantially differ from one author to the next. The contours chosen here have substantially taken into account those represented in BRULHET, PETIT-MAIRE *et al.* (1999).

Periglacial features are also known in non-glaciated mountains such as basement uplands of Western and South-Western Europe, Mediterranean mountains, eastern Anatolia mountains above 2700 m and the Elburz mountains above 2100 to 2300 m (VELICHKO & NECHAEV, 1992; WILLIS, 1994).

III.2.2.2.- Eurasia

Conditions all across Northern Eurasia appear to have been dry and treeless, dominated by polar desert or semi-desertic steppe-tundra. The buried soils show chemical and morphological indicators of arid conditions, together with other "desert" features such as windblown sands, wind-sculpted pebbles, wind-eroded hollows and the sediments of intermittent desert lakes. The very lack of animal fossils or organic sedimentation from most of Northern Eurasia during this period further suggests aridity.

Cold and arid conditions also prevailed in Central Asia, where oscillations in lake levels are correlated with influxes of meltwater from the northern ice sheets (Caspian Sea) or from the Pamir glaciers (Aral Sea). In contrast, more variable conditions may have occurred in the mountain areas of Iran and Anatolia. In eastern and central Turkey, high lake levels at the LGM - 18 ka - existed at the same time as arid steppe-like vegetation (LANDMANN *et al.,* 1997; ALSHARHAM *et al.,* 1998), perhaps due to storms occurring mainly during the winter season (PRENTICE *et al.*, 1993; ROBERTS *et al.*, 1999).

III.2.2.3.- Central and Western Europe

Since 31 ka up to just before 13 ka B.P., the last part of Glacial period had been very cold and dry throughout Europe, the main ice build-up having occurred in the Alps between 24 and 21 ka B.P. (VAN HUSEN, 1996). Ice caps covered the Alps and the Pyrenees. The maximum extent of mountain glaciers outside the main ice sheets (Alpes: 250 000 km^2) during Late Pleistocene is mostly related to the LGM, except, at least, for the Pyrenees, where it is older (ca 38.4 ka B.P.: BORDONAU *et al.,* 1993) and where the LGM presents a smaller extent. In the Alps, the Equilibrium Line Altitude (ELA) was 1200 m lower than nowadays. The Western Alps were more heavily glaciated than the Eastern Alps, where big glacier tongues flowed along the main trunk valleys (Enns, Mur, Drau) and could build piedmont lobes like in the western Alps.

Small ice caps, piedmont or valley glaciers covered the mountains of Central Europe (Tatras, Carpathians) and smaller mountains of France (the Vosges: SERET *et al.*, 1990; the Massif Central: VAN VLIET-LANOË *et al.*, 1991; Corsica: CONCHON, 1977), Italy (JAURAND, 1998) and Spain (PÈREZ-OBIOL & JULIA, 1994). The precise age and extent of local glaciers at the LGM may be prudent in order to dismiss the concept of a single LGM maximum for Isotope Stage 2, as far as local glaciers are concerned. Rapid glacier build-up at the LGM in the Eastern Alps is reported to have occurred after ca. 24 ka B.P. (FOLLIERI *et al.*, 1988 and 1993; VAN HUSEN, 1997; CLAPPERTON, 1999). There was also dissymmetry of extent between the western and eastern parts (i.e. Vosges, Alps, etc.) due to the North Atlantic westerlies situation, or between northern and southern sides (i.e. Pyrenees; HÉRAIL *et al.,* 1986).

All parts of Europe were much colder than today, with the greatest cooling in winter. Around the latitude of South-Central Germany and North-Western Ukraine, FRENZEL *et al.* (1992) present maps suggesting August mean temperatures of about 10-11 °C, comparable with much of the northern coast of Siberia at present. Winter (February) mean temperatures were at least as low as -19°C in Central Germany and -27°C across most of the Ukraine. In Southern Europe, across most of the Mediterranean zone, temperatures were perhaps 8-10 °C lower than nowadays in both summer and winter (FRENZEL *et al.*, 1992; GUIOT *et al.*, 1992 and 1993). Drifting sand and wind erosion were common in North and Central Europe.

Actually, the ice cap and North Atlantic sediment chronology suggests that the maximum cold and aridity of this general period is a composite picture of two large cooling phases or "Heinrich events" (around 21 and 17-15 ka cal. B.P.). These may have been separated by a somewhat milder period, lasting a few thousand years, which shows up in southern European pollen records as a "blip" of pine (*Pinus*) pollen (HUNTLEY, 1992; HUNTLEY & PRENTICE, 1993).

III.2.2.4.- Atlantic

a.- The periglacial rim. The near Atlantic subpolar ocean is a narrow zone, between 55° and 45°N, subject to (1) severe conditions of atmospheric gradients: winds (tracks of strong depressions) and temperature (winter air <10-16°C); (2) major fluctuations of seasonal sea ice, from Ireland to the western approaches of Channel (see the oblique south-western border of winter ice, in geographic connection with the permafrost limit); (3) latitudinal descent of icebergs; (4) low salinity conditions (< 35‰) and (5) annual variations of sea surface temperature increasing southward, i.e. in August, 2°C west of Ireland, 6-7°C offshore Brittany (CLIMAP, 1976; MCINTYRE, KIPP *et al.*, 1976; LABEYRIE *et al.,* 1992; HARRISON & DIGERFELDT, 1993); it is an annual amplitude probably similar to the present conditions.

b.- The humid Atlantic Westerlies zone: Approximately south of 47°-45°N, the prevailing winds gradually turn to the west. The conditions are deeply different from the present-day ones. Over the Eastern Atlantic (southern Bay of Biscay and Iberian - Moroccan seas), the driving forces of the ocean give more importance to the eastward drift. The results are: (1) a light warming of the SST, mainly south of 42°N with a marked thermal

gradient visualised on the map by the hydrological North Atlantic Polar Front. The August temperature is 10°C (southern Bay of Biscay) to 18°C (Morocco); in February, respectively 2°C and 12°C (annual variation: 8 to 6°C); (2) a notable increase of salinity, up to 35.3‰ (south of polar front, in the Iberian waters) in relation with the evaporation (winds, agitation and heat content); (3) an equatorial displacement of the North Atlantic Drift. The hydrological deep structure is probably significantly different from the modern one: a reduced role for the production of Intermediate Arctic Water and, by compensation effect, the dominant volume and action of Antarctic Bottom Water.

III.2.2.5.- Mediterranean basin

a.- The Western Mediterranean basin underwent a dry and cold form of continental climate, mainly in the vicinity of the glaciated Alps and of plains with discontinuous permafrost or severe winter frost. But the general environment differs little from the present, with the same counter-clockwise gyrations as today, in spite of a marked heat deficit (3-4°C colder than today). The Mediterranean was still exporting dense but considerably reduced outflow water to the Atlantic during the last pleniglacial low sea-level (BÉTHOUX, 1993; THUNELL & WILLIAMS, 1989; VERGNAUD-GRAZZINI *et al.*, 1989; BIGG, 1994; TURON *et al.*, 2000).

b.- The Eastern Mediterranean (mainly the Ionian and Levantine Seas) presents a similar elongate gyre pattern, but reveals several differences and deficiencies. (1) A steep and quasi-permanent temperature gradient has dominated the connection between the west and east Mediterranean basins (difference: 5°C in February, 3°C in August), and with the northern small semi-enclosed basins (Otranto basin, Cretan Sea) in contact with cold terrestrial runoff from the inter-mountain continental shelves (Adriatic and Aegean Seas). Their influx of cool and fresh waters fluctuated from 10-11°C (February) to 15-17°C (August). (2) As regards the environment of the elongate deep basin, the budgets were (a) highly positive for temperature: the SST resembled the modern thermal regime with 6-10°C more in the western basins, (b) deeply negative for the water balance because of southern continental aridity, runoff diminution and strong evaporation. The superficial salinity was 41.5‰. The Cyprus basin was, as today, an area of dense-water formation, but the dense, oxygenated, and nutrient-enriched waters of the Otranto and Cretan basins described protruding incursions into the deep Levantine basin. Consequently, it was a specific basin with increased temperature, salinity, oxygenation and stability (ROSSIGNOL-STRICK *et al.*, 1982; ARKHIPOV *et al.*, 1995).

III.2.2.6.- Enclosed seas or lakes

a.- The Black Sea. In spite of abundant input of meltwater from the Scandinavian ice sheet through the Dniepr valley, its level (-140 or -150 m) dropped considerably lower than that of the Bosphorus and Marmara Sea sills and thus transforming into an open lake whose waters became increasingly fresh and cold (winter: 2°C; summer: 13-15°C; today, respectively, 6-8°C and 25-28°C). It is the "New Euxine" freshwater phase of the regional eustatic evolution.

b.- The Caspian Sea possessed a markedly positive hydrological budget since it received abundant water from proglacial lakes drained by the Volga river. It became the greatest lake in the world (Early Khvalynian transgression). The area of this briny, euxinic and cold lake reached approximately 600 000 km^2, but it was subject to strong short term fluctuations. Though the LGM period roughly coincides with the Yenotavian regression (-50 m; 24-17 ka), the map presents one of the more representative higher sea-levels which prevailed during the latest parts of the Late Pleistocene. Drainage towards the Azov - Pontic Basin through the wide and continuing uplifting Manytsh Strait could occur during periods of high level stasis but it is not ascertained that it could be maintained during the event represented here (Late Khvalinian transgression: 0-5 m; 17-8 ka; MAMEDOV, 1997).

c.- The Aral Sea was also characterised by a high level, possibly related to abundant input from the Mansi palaeolake formed in Western Siberia along the margin of the Kara ice dome and from the Amou - Daria. Possible spillways could be established towards the Caspian Sea during the highest seastands (VELICHKO, 1984; LÉTOLLE & MAINGUET, 1994; EHLERS, 1996).

III.2.2.7.- North Africa, Arabia and Middle East

Aridity, and to a lesser extent cooling, are the major features of Africa during the full glacial period. Ancient sand dune distributions suggest that the Sahara desert zone extended hundreds of kilometres further south than at present, compressing the other vegetation and climate zones equatorwards (SARNTHEIN *et al.*, 1982; LÉZINE, 1989; LÉZINE & CASSANOVA, 1989; THORP, 1994; TCHAKERIAN, 1994). However, in a restricted area of the North-West Sahara, greater winter rainfall seems to have given rise to moister conditions, with a belt of semi-desert appearing to the south of the present-day desert margin until *ca* 20 ka B.P. (HOOGHIEMSTRA *et al.*, 1992).

There are indications from various parts of Africa (Central and Eastern Africa) and the Arabian peninsula that maximum aridity may have occurred slightly after the global LGM, at around 17-15 ka B.P., although the LGM itself seems to have been generally much more arid than present conditions (GASSE *et al.*, 1990; EL-NAKHAL, 1993; GASSE & VAN CAMPO, 1994; ADAMSON *et al.*, 1995; HEATHCOTE, 1998). Plankton indicators of upwelling intensity off the coast of Somalia suggest that the summer monsoon (which supplies much of northern, East and Central Africa with rainfall) may have been weaker than today, and reached its weakest at around 15.8-12.5 ka B.P. (17.8-13.8 ka cal. B.P; ZONNEVELD *et al.*, 1997; CAULET *et al.*, 1992; SIROCKO *et al.*, 1993; LÉZINE & VERGNAUD-GRAZZINI, 1994).

III.2.2.7.- Arid and upwelling oceanic systems

South of 35°N, during the Last Glacial Maximum, the vast zone running from the Central Atlantic to the North-Western Indian Oceans, was influenced by the trade wind regime (LAMB & WOODROFFE, 1970).

a.- The Saharan continental margin. The dominant feature is an intensification of the trade wind belt and the oceanic advance of arid conditions. On the East Atlantic sector and African seaboard, a quasi-permanent low pressure interval was established, as today, between the Central Atlantic and African anticyclones. The northeast prevailing winds, blowing towards the ITCZ (Intertropical Convergence zone), were colder than present and permanent. There is, however, a marked winter reinforcement (CLIMAP, 1976; McINTYRE, KIPP *et al.,* 1976; COHMAP, 1988; HOOGHIEMSTRA *et al.*, 1992; EL FOUKALI, 1995). The wind-driven Canary Current was the Eastern Boundary Current of the subtropical gyre.

b.- The Indian Ocean Corner. The trade wind regime was, as it is today, dominated by the reversion northeast (winter)-southwest (summer) flux system named monsoon. Throughout the year, the prevailing winds were dry and weak. In contrast with the Saharan system, the decrease of the monsoon cycle seems due to the reduction of land-sea temperature and pressure contrast, particularly in summer, when the increase in thickness and/or areal extent of the Asian snow cover reduced the gradient driving the oceanic flux. (1) On the Persian Gulf, Red Sea, and the Gulfs of Aden and Oman, the wind blowing from dry continental areas exerted a strong action on the semi-enclosed water masses: high salinity (surface and bottom) and reduction of ventilation. In particular, the low sea level nearly isolated the Red Sea from the Indian Ocean, and developed conditions capable of lowering oxygen content and reducing the ventilation in the axial through. (2) On exposed oceanic areas, the Somali and South Arabian margins, the winter trade winds, blowing across the Arabian Sea, drove a slow and warm expanding water mass. In return, under the influence of summer monsoon winds, Ekman pumping associated with the westerlies created two large regions of upwelling (SUS and AUS on the map, for Somali and Arabian Upwelling Systems). These upwelling belts were characterised by a relatively cool (25°C) and nutritive surface (HAQ & MILLIMAN, 1984; ANDERSON & PRELL, 1991, 1993; ANDERSON *et al.*, 1992; Sirocko).

IV.- DESCRIPTION OF LGM DOMAINS

IV.1.- Sea, coasts and land

During the maximum negative stage (-120 to – 125 m), the land surface was considerably larger than today. Streams and glaciers were shaping vast areas of now submerged continental shelves, particularly in Europe where about 4.10^6 km^2 were exposed, nearly doubling its present-day area (e.g., the emerged floors of the North Sea, the Channel or the wide emerged shelves of the Bay of Biscay, of the Northern Adriatic Sea, etc.). Over 200 000 km^2 emerged in the Persian Gulf, while the Red Sea became a nearly closed hypersaline basin (HAQ & MILLIMAN, 1984). The major consequences were (a) the extension of continental and dry climates, (b) the reduction of epicontinental sea biotopes, (c) the decrease of oceanic biodiversity, and (d) the lowering of the CCD.

Areas of non-deposition were submitted to erosional processes whose distribution corresponds to the LGM palaeoclimatic zonation (FRENZEL *et al.*, 1992): (1) periglacial processes north of the Mediterranean sea and in non-glaciated mid-latitude or even subtropical mountains (frost-shattering, soil disturbances in relation with freeze-thaw cycles, slope mass-wasting, solifluxion, etc.); (2) eolian processes, sheet or rill wash and mass movements in cold and steppic environments; (3) intense chemical weathering and pedogenesis in more restricted domains of mild temperate climate, below the open forests that remained around the Mediterranean Sea (southern tips of the peninsulas); (4) reduced rates of mechanical or salt weathering in and around the deserts (except in the more humid mountains and piedmonts of NW Africa).

Regoliths and colluvial mantles of debris reworking older deposits (loess, alluvium, etc.) as well as frost shattering products are the most typical products related with the Late Pleistocene periglacial conditions (e.g., LAUTRIDOU, 1985; BALLANTYNE & HARRIS, 1994).

IV.2.- Eolian

North of the Mediterranean Sea, an important eolian activity takes place under dry and cold climatic conditions, in tundra-type or steppic environments. The largest areal extension of loesses and loess-like sediment is related to the Last Glaciation (PÉCSI, 1992); their lithostratigraphy is highly diversified from one region to another and reflects the sequences of humid and dry oscillations (sedimentation gaps, palaeosols). Two kinds of deposits are represented: loess and coversands. The loess particles were transported in suspension and deposited in all kinds of sediment traps, such as basins, dells, foot-slopes, plateau surfaces and basins. All transitions are found between loess and coversands (PÉCSI, 1992; EHLERS, 1996; LOWE & WALKER, 1997). This is the reason why they are not separated on the map.

Loess or eolian silt deposits are mainly formed of quartz grains. They also contain some carbonates (up to 40%), feldspar grains, clay minerals and micas. A minor component of heavy mineral may be used to recognise the origin of the deposits. Together with other eolian deposits, they are widely spread in Europe and in central Asia. They form continuous and locally thick covers (4 to 10 m) in Northern France, Belgium and the Netherlands, whereas they are more discontinuous in

Southern England or in the Jersey Island (LAUTRIDOU, 1985; ANTOINE *et al.*, 1998). The largest accumulations are known in Central and Eastern Europe, north of the main mountain ranges, in the Middle Danubian Basin, in Ukraine, in the Russian plain and in Central Asia. Most of them are reworked from glacial and glacifluvial deposits and from the wide braided channels in alluvial valleys, swept by the wind during the dry seasons. In Northern France, eolian silt also originated from adjacent emerged sea floor (North Sea, Channel). Other deposits were related to large alluvial valleys (Rhone valley, Rhine graben, etc.).

Loessic sands were also deposited around the deserts, south of the Mediterranean Sea (Tunisia) and in the Middle East. The geographical area covered by mobile sands was roughly the same as nowadays. Nevertheless, it is suggested that a somewhat more humid climate characterised the North-Western Sahara, resulting in steppic environments more favourable to loess deposition than to sand migration (ROGNON, 1990; LIOUBIMTSEVA, 1995).

IV.3.- Fluvial and lacustrine domains and deposits

IV.3.1.- Drainage patterns and lakes

The main differences between the LGM drainage patterns and the current hydrography are found on emerged continental shelves, along glacial margins and in the deserts.

As the northern parts of the North Germany and Poland plains were covered by the Fenno-Scandinavian ice sheet, rivers from mountains of Central Europe and meltwater channels were forced to flow westwards along the morainic ridges of the glacial margin. Hence, the lower Elbe valley, between the Havel river and the emerged floor of the North Sea was the main outlet. To the east, four or five major channels were formed along the moving margin of the Weichselian ice sheet; they are called Urstromtäler or pradolina (EHLERS, 1996). Several proglacial lakes were formed along this margin in Russia. Excess water could flow southwards along the Dniepr and Volga valleys, respectively to the Black Sea and the Caspian (ARKHIPOV *et al.*, 1995).

The southern North Sea (Southern Bight) was drained southwards through the Pas-de-Calais, to the valleys of the Central Channel and into the Manche lake, so that the London Basin belonged to the same catchment area as the Paris Basin. This system was ended by an enormous deep-sea fan. Increasing production of nannofossils during the period of low sea level suggest that the hydrologic regime favoured nutrient inputs as well as nutrient recycling (LE RICOLAIS *et al.*, 1998; AUFFRET *et al.*, 2000).

South of the Mediterranean Sea, dry conditions prevailed, especially south of the Maghreb. The White Nile was obstructed by mobile sands, while inputs from the Blue Nile and Ethiopian tributaries were becoming irregular, so that the main trunk valley was rather similar to a large wadi and could be invaded by dune fields (BUTZER, 1980; ADAMSON *et al.*, 1995). In the Afar depression, the Lake Abhe was completely dry at 17 ka, after a high level which had occurred at about 21-22 ka B.P. (GASSE, 1977). The situation seems to be more complicated in the Middle East, since high levels are known in the Dead Sea between 18 and 12 ka B.P. and fluctuating lake levels are described in lake systems of Central Anatolia until 17.5 ka B.P. (Konya: KARABIYIKOGLU *et al.*, 1999).

Lakes are reported within most of the area mapped here. The biggest of them are tectonic lakes and sebkhas (Caspian and Aral Seas, Dead Sea, lakes of Anatolia, Tunisian chotts, etc.), whereas volcanic lakes are much smaller. Glacial lakes cover large areas in Northeastern Europe, along the ice sheet margin.

IV.3.2.- Deposits and environments

Type (a): the periglacial plains. Western European shelves are characterised by gentle undulating spreading out of sand/gravel-veenered flats (1) created by lateral deposition of braided streams, reworked by periodic solifluction (see southern limit of permafrost); and (2) undercut by axial denudation along the course of main collecting rivers (ex. the central Channel deep in the bed of the Seine River). More generally, seasonal flooding due to snowmelt, and increases in sediment yield due to combination of a sparse vegetation cover, strong frost shattering and ground disturbance by periglacial processes resulted in wide systems of braided channels, now preserved as fluvial sand and gravel terraces (VANDENBERGHE *et al.*, 1994; LOWE & WALKER, 1997). Such sediments are now buried in areas affected by active downwarping (Netherlands, eastern Romanian plain).

Type (b): alluvial wadi plains and terraces (ex. Iberian, African, and Arabian Sea borders, non-glaciated mountains of the Tethyan domain, deserts). In front of open sea, more or less extensive infillings or aprons of torrential material (coarse, poorly sorted rubble, sand and gravel), are deposited in response to rain storm discharges. By places (mainly in the southern domain), the eolian deposition can take the forms of wide sheets of dunar sands, bordered along the coast through large accumulative chains and ridges, more or less lithified (CASTAING *et al.*, 1972).

Type (c): Carbonate-cemented plains and platforms (e.g., Mediterranean shelves, Red Sea, Persian Gulf). The fluvio-eolian deposits in the river valleys, sebkhas and terraces (coarse to fine sand and pebbles) are submitted to leaching, calcified and, locally, shaped in platforms bounded by fluvial scarps and coastal cliffs. Around the Red Sea and Persian Gulf, the inherited coral reefs emerge in isolated rises (THIEDE & SUESS, 1983). In the present deserts, enhanced aridity resulted in phases of aggradation or obstruction of valleys by mobile sands.

IV.4.- Glacial and glacifluvial areas

Adjacent to the LGM ice front, the pre-existing glacial morphology (ground moraines) of North European shelves and emerged plains is (1) masked and reworked by plain deposits, (2) shaped by proglacial streams (e.g., St-George Channel), or (3) incised in narrow deeps (e.g., Devil's Hole, North Sea). Well-developed moraines are found in Denmark, Northern Germany, and in the lowest areas of Poland, where they were built by glacial lobes. In other areas, the ice margin is only marked by hummocky moraines (Lithuania) or unclear glacial or glacifluvial accumulations of sand, gravels and silt. Similar sequences of sediments were deposited around the main glaciated mountains whereas thick outwash deposits ultimately preserved as terrace systems were laid down in downstream river valleys (Rhône, Durance, Rhine, Donau and other piedmont valleys in Eastern Europe and Central Asia (EHLERS *et al.*, 1991; EHLERS, 1996; VAN HUSEN, 1997).

IV.5.- Shallow environments

A wide variation of salinity is the prevailing character of the following almost enclosed regions: (1) lagoonal-fluvial mouth settled with lutite (depressions) and fine sands (dunes, barrier islands) (e.g., Armorican shelf, Grande Vasière); (2) semi-enclosed shallow depression and basin subject to a voluminous fluvial supply (e.g., southern Adriatic shelf and Otranto basin); (3) evaporitic and shallow bioherm platforms with narrow lagoonal salt mud, barrier coral reefs and beachrock.

IV.6.- Coastal-shallow marine belts

The regions mapped here correspond to:

- the present-day outer continental shelf submitted during the LGM to a detrital and removal coastal-fluvial supply: the offshore Scotland to Aquitanian margin, the outer Golfe du Lion and Golfe de Gabès, the clastic muddy offshore Nilotic delta. The wide southern Celtic shelf is the most significant domain with its subparallel, symmetric and *en échelon* sand ridges shaped by strong tidal currents (BOUYSSE *et al.*, 1976). Terrigenous and calcareous fluxes increase with decreasing sea level (AUFFRET *et al.*, 2000);

- the high productivity zone. The tropical African-Arabian upper slope (mainly between 1000-1500 m), adjacent to the LGM arid zone is the locus of an abundant biodetrital (foraminiferal, diatomaceous) ooze, associated with eolian supply. A distinctive assemblage of superficial water plankton species is deposited and preserved in the sediment, creating in the sedimentary column a geologic record of the coastal upwelling and terrestrial wind activity. The Saharan margin is the centre of maximal organic-matter accumulation. By its nature and its thickness, the depocentre of the higher bioproduction of the ocean reflects eolian impact on the sea (intensity of upwelling, fertilising action of the aerosols) and eolian input from the land (velocity of the trade winds, volume and composition of the dust plume flux). From the point of view of carbon budgets during the LGM, the Saharan slope is the upper and the main carbon reservoir (see below). The depocentre played a similar role to the present continental shelf through a downward transfert of function (SARNTHEIN *et al.*, 1988; SARNTHEIN & WINN, 1990; BERTRAND, *et al.*, 1996; GROUSSET *et al.*, 1996). In the Indian Ocean, the Oman and Somali homologue provinces differ from the Saharan example. The less intense and only summer upwelling reduced in space and time the sedimentary response of the water column (low rate of biogenic accumulation; SIROCKO, 1991; SIROCKO *et al.*, 1991 and 1993; ANDERSON & PRELL, 1993). The immediate post-LGM period (near 17-16 ka B.P.) is marked by an abrupt and drastic sedimentary change, correlated with the beginning of decrease of eolian dust Saharan sedimentation (SARNTHEIN *et al.*, 1982).

IV.7.- Deep marine

This large and deep area of dominant hemipelagic biogenic ooze (to 5000 m depth) presents a latitudinal and regional diversity induced by lithic supply. From north to south, and west to east:

Type (a): the Atlantic northern range (north of polar front) gives the last image before the final northward transgression of the North Atlantic Drift. The region is clearly defined by (a) the penultimate deposition phase of the "Ruddiman belt" (known as "Heinrich Layer 2"): the ice rafted input unloaded in a dense (down to 45°N) or scattered (southward) carpet debris; (b) a slowly removed muddy veneer (planktonic foraminifera of Arctic origin) in a confined environment as expressed by an oxygen depletion of deep water. The causes are the absence of Norwegian bottom water and the sluggish progression of the North Atlantic Deep Water (residence time about 4 times that of today) (e.g., DUPLESSY, 1982; MCINTYRE, KIPP *et al.*, 1976; LONSDALE, 1982; RUDDIMAN, *et al.*, 1989; LABEYRIE *et al.*, 1992; LEHMAN & KEIGWIN, 1992; CACHO *et al.*, 1999; TURON *et al.*, 2000).

Type (b): the polar-front upwelling-belt interval. The detrital supply (shale and clay) originates from the margins by gravity processes along the canyon axis (levees and deep-sea fans), continental rises (furrows, mudwaves, slides and slumps) and abyssal plain floor. The intervention of contour current (Antarctic bottom water) in still problematic.

Type (c): the Tropical edge. The abyssal ooze zone was: (1) submitted to a partial and distal flux of biogenic particulate matter carried down from the upper high-productivity depocentre. Because of this spatial continuum with the continental slope, the Saharan ooze zone can be considered as the same sink of carbon; (2) contaminated by eolian particles transported by wide Saharan dust plumes. The lithic fluxes were 2 to 4 times higher than today, enhanced mostly during the winter (SARNTHEIN *et al.*, 1982; THIEDE *et al.*, 1982; BERTRAND

et al., 1996; MARTINEZ *et al.*, 1996; GROUSSET *et al.*, 1998).

Type (d): The Mediterranean. By its configuration (virtual isolation from Atlantic and Black Sea), the Mediterranean was a sedimentary trap where the calcareous pelagic oozes (foraminife - pteropod assemblages in response to annual variations in surface water temperature) were influenced by the terrigenous and water supplies from a surrounding landscape colder and drier than today. During the LGM, the semi-enclosed basins suffered a radical depositional change: (1) end of falling sea-levels, well-mixed water column and high-accumulative conditions, mainly in western basins (deep-sea fans, last turbidite abyssal layer, etc.); (2) beginning of Holocene s-l. elevation, lowering sedimentation rate, slackening (locally stoppage) of bottom circulation and dense-water formation, mostly in eastern basins (stagnation and sapropel formation by episodic Holocene freshwater inputs) (TURON *et al.*, 2000).

Type (e): The Indian corner (offshore Oman and Somalia). The reduced accumulation rate of biogenic deposits (Diatoms and Radiolaria) and the increased eolian fluxes are the signals of a weak summer upwelling originated by the westerlies repressed by the convergence with the monsoonal system.

IV.8.- Deeper carbonate ooze

Very low sedimentation rate off the westerlies area and below the 5000 m isobath, at the vicinity of CCD (calcium carbonate compensation depth).

IV.9.- Hypersaline

Sebkhas. The sebkhas represented on the map are mainly those that can be observed nowadays. Since most deserts were drier than presently, it is not ascertained whether all of them were receiving deposits during the LGM. In Northern Africa, the palaeolake Djerid underwent a marked regression, which reduced it to shallow sebkhas at about 18 ka B.P. (ROGNON, 1990; GASSE & VAN CAMPO, 1994; PETIT-MAIRE *et al.*, 1994; ALSKARHAN *et al.*, 1998).

Central Red Sea

The corresponding deposit is a biogenic ooze trapped and preserved in a nearly isolated, little ventilated high salinity basin (ALMOGI-LABIN *et al.*, 1991).

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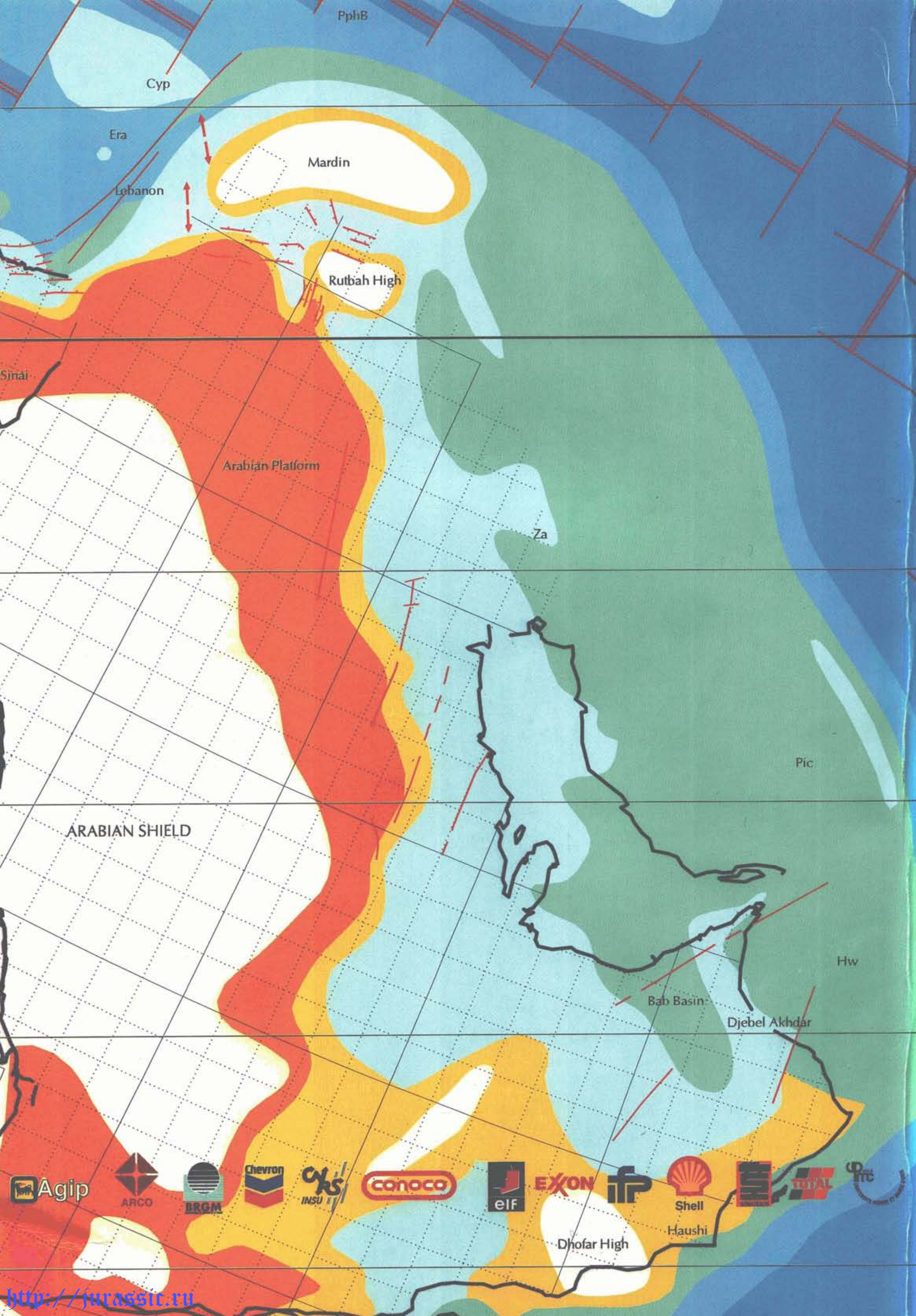
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